

DEBRIS FLOWS AND SLUSH AVALANCHES IN NORTHERN SWEDISH LAPPLAND

Distribution and geomorphological significance

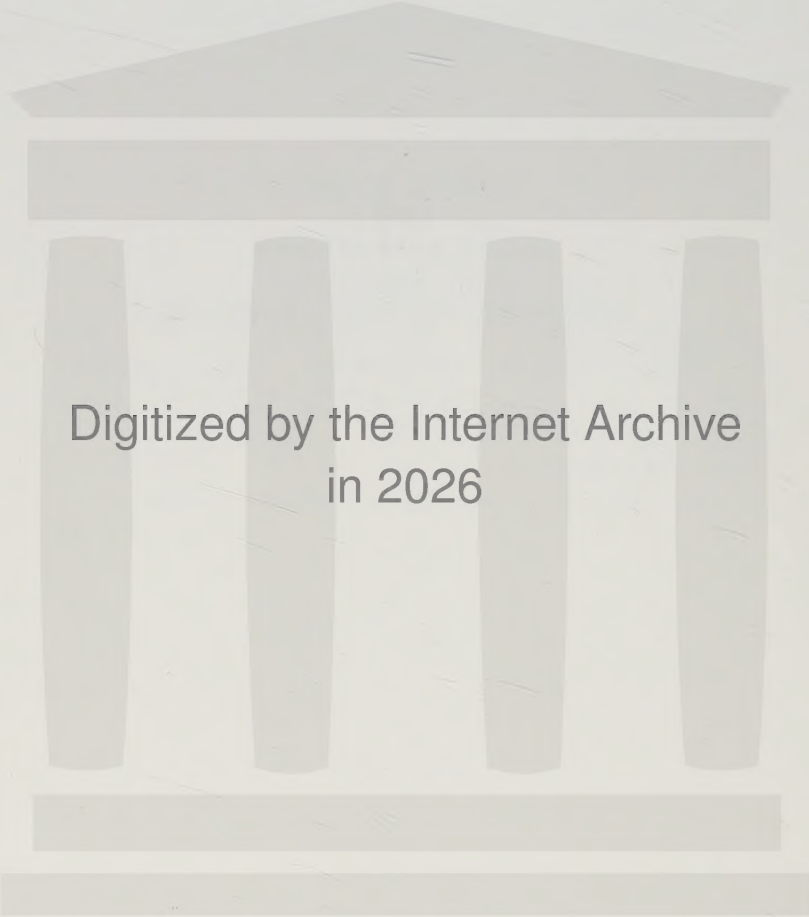
ROLF NYBERG



MEDDELANDEN
FRÅN LUNDS UNIVERSITETS
GEOGRAFISKA INSTITUTION
AVHANDLINGAR XCVII

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AVHANDLINGAR XCVII

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Cover: Slush avalanche track at Kaskarieppe, Nissunvagge. Photo 16.7.1982.

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Foreword

The studies concerning geomorphological processes and landforms presented in this thesis were initiated in 1977 and have been in progress with summer field work in northern Sweden up till 1983, supplemented by winter season field studies during the years 1980-83. Some additional observations were made in the following years in connection with further field work in the area. Except for two and a half years of research scholarship, I have combined work on the thesis with a part-time post at the Department of Physical Geography in Lund. My interest in geomorphology originates largely from my first teachers, Dr. Åke Hillefors, formerly in Lund, and the former head of the Department, Prof. Karl-Erik Bergsten. The thesis was envisioned while partaking in a field course on slope processes in the Abisko area in 1976 under the leadership of Dr. Lennart Strömquist, Uppsala University, who during the course and on several occasions later has stimulated my interest in the subject. In the same year, our department came under the directorship of Prof. Anders Rapp, who with his personal vast experience of mass movements in the Swedish and other mountains became my supervisor and interested me in continuing some of his earlier studies in Swedish Lapland. I have had the

great advantage of being able to draw upon his experience in my investigation, both at the department in Lund, and perhaps most valuably, during several joint field visits in the Abisko mountains.

Much of the field work was made living in a tent and was considerably facilitated by the access to the Abisko Scientific Research Station as a 'base camp'. The personnel at the research stations at Abisko and Tarfala (in the Kebnekaise mountains) have been both helpful and hospitable during my stays there. In particular thanks are due to Prof. Mats Sonesson, Superintendent Nils-Åke Andersson, Dr. Björn Holmgren and technicians Anders Eriksson and Bengt Vanhatalo at Abisko and the late Prof. Valter Schytt at the Tarfala Glaciological Research Station. Prof. Wibjörn Karlén, Stockholm, gave useful advice concerning the study of snow conditions at avalanche sites. I am also indebted to Kungliga Vetenskapsakademien for granting the use of their hut in Kärkevagge. During the field work, I have received valuable assistance by Anders Ekström, Bengt Gillenius, Kerstin Melander, Yvonne Pålsson, Jan Rapp, Jöran Rehn and in particular Ann-Margreth Berggren and Lars Bärning. However, most of the work was done by myself working alone, which of course to some degree has limited the amount and type of field work done.

Colleagues at the Geography Department in Lund, especially Dr. Bo Malmström and Dr. Owe Palmér, provided valuable input of ideas in the planning and execution stages of the work. Am. Herbert Blond set an example as a geographic scholar of admirable versatility and perseverance. Special thanks to FK Mikael Stern, who afforded time-saving help with adjustments of the instrument during photogrammetric mapping and to FK Lennart Olsson and FK Lars Bärning, who guided me on the very first step towards enlightenment in computer religion. Kristoffer Oliverson gave valuable help in the translation from a book in Icelandic.

The C-14 and wood laboratories at the Department of Quaternary Biology/Geology at Lund University have helped with datings of soil and wood samples, and thanks are due to Dr. Sören Håkansson and Dr. Thomas Bartholin. I am also grateful to State Meteorologist Erik Schmacke at SMHI for granting the loan of a snow penetrometer. The photographic work was carefully done by photographer Reszo Laszlo, assisted by FK Ulrik Mårtensson, who made the best possible of my originals taken under unfavourable weather conditions in general. Photographs are by myself except when otherwise stated. A large number of oblique air photos were kindly put at my disposal by Nils-Åke Andersson, Abisko Research Station, and have been of particular value as a basis for parts of the work. A few of these photos are reproduced in the study. Assistance with the drawing of some of the Figures was provided by Klara Laszlo, Eva Särbring and Inger Lander. Permission to publish maps and air photographs was obtained from Statens Lantmäteriverk 1985-09-25. Financial support for the study was received from Naturvetenskapliga Forskningsrådet (NFR), Svenska Sällskapet för Antropologi och Geografi (SSAG) and Abisko Naturvetenskapliga Station. A travel grant was received from Nordiska Samarbetsfonden for a study visit to the Jotunheimen area in Norway.

Professor Anders Rapp has read the manuscript and made valuable suggestions on improvements. Dr. Bengt Orestrom checked the language. To the persons and institutions mentioned above, to the personnel at the Department of Geography in Lund, to many others who in different ways helped me, and to my parents, who supported me all through the work, I extend my warmest thanks.

Rolf Nyberg
Lund, October, 1985

1. Introduction

1.1 Studies of mass wasting in geomorphology

Slopes and the processes acting on them is one of the central concerns of geomorphology, the study of landforms and their development. Although seemingly unchanging for long periods of time, slopes undergo an evolution as material is transferred from higher to lower ground. This cascade of slope debris, recently often viewed in terms of energy (Chorley & Kennedy 1971, Caine 1976), is ultimately governed by the influence of gravity and solar radiation. A long-term trend of relief reduction hence affects the landscape unless diastrophic processes intervene. The nature of slope development over long-time periods has been at the core of the different models put forward to explain landscape evolution, see e.g. Young (1972), Carson & Kirkby (1972) for reviews. Broadly speaking, slope evolution may proceed in two ways, either in slow, more or less continuous changes, or in rapid stepwise adjustments to imposed environmental stresses. The case is particularly clear concerning movements of material downslope mainly by gravity alone. Traditionally, a division has been made into slow mass movements, for instance soil creep, and rapid mass movements, often of a catastrophic nature, such as rock slides (Sharpe 1938).

Statham (1977, p.70) makes a distinction between mass movements and other sediment transport processes (flow processes) on the basis of movement of large aggregates of grains by gravity alone versus movement of mainly single grains dispersed through a fluid medium, functioning as a transporting agent. In reality this distinction is often blurred by gradual transitions between processes (Op. cit). This is particularly obvious regarding two forms of rapid mass movements dealt with in the present study. In debris flows, water is involved in the movement to a high degree. The same applies to avalanches of wet snow (slush), with some of the water in solid form.

Snow avalanches are mass movements strictly only when the displacement of the snow is considered. In a geomorphic context, where interest lies in the erosion and transport of rock and soil debris by the processes, they could equally be regarded as flow processes, with snow as the transporting medium. In many classifications of mass movements, the processes involving snow are left out, e.g. Sharpe (1938), Varnes (1958), Carson & Kirkby (1972) and Carson (1976). Selby (1982) includes debris flows both under 'mass wasting of soils' and under 'water on rock slopes', but snow and slush avalanches are only treated under the latter heading. Embleton & King (1975), Brunsden (1979) and Finlayson & Statham (1980) exclude avalanches involving snow in their treatment of mass movements. However, Rapp (1960b), Nobles (1966) and Young (1972) classify them under rapid mass movements, as will also be done in the present study. More generally, the geomorphic importance of snow and slush avalanches is frequently overlooked even in recent textbooks in geomorphology (e.g. Bloom 1978, Chorley et al. 1984).

Snow avalanches have been widely studied for practical reasons (see e.g. Perla & Martinelli 1975 for a review), but to a much smaller extent with respect to their role in mass wasting on slopes (cf. Luckman 1977). Similarly, because of the obvious practical importance of studying slope stability and because in many countries mass movements are a source of soil erosion on cultivated land, considerable work on mass movements has been done by engineers and agricultural researchers. These contributions are well covered in recent text- and handbooks on slopes and general techniques in geomorphology, which give ample space to the geotechnical and soil erosion aspects (e.g. Young 1972, Carson & Kirkby 1972, Embleton & Thornes 1979, Goudie 1981, Selby 1982, Gardiner & Dackombe 1983). There has also been a trend towards an emphasis on the applied aspects of the discipline (e.g. Cooke & Doornkamp 1974).

Apart from the practical value of slope studies, a theoretical approach with regard to forms and processes of natural slopes can easily be justified. A knowledge of the natural rates of geomorphic processes in different areas is essential as a basis for the appreciation of the human impact on the environment. Furthermore, information gained from studies of landforms can be used to shed light on earlier climates and climatic variations provided that the climatic control on landform development can be assessed (cf. Büdel 1977).

1.2 Purpose and approach of the study

The study of mass movements assumes particular importance within the subfield of mountain geomorphology (Hewitt 1972, Barsch & Caine 1984), due to the interaction of high relief and active climatic factors in high mountains. The present thesis deals with mass movements occurring in a periglacial mountainous environment of northern Sweden. These phenomena are of interest not the least owing to their potential hazard to increasing scores of visitors. The relative importance of different mass movements during the recent climatic conditions and generally in postglacial time in Swedish Lappland is a subject that was investigated in a major work by Rapp (1960b), who studied the recent rates of slope processes in Kärkevagge trough valley west of Abisko. The present study, although with a partly different approach, concentrating on two of the several recently active processes (debris flows and slush avalanches) and involving more of regional outlooks, represents an attempt on the basis of the work of Rapp and others to further extend our knowledge of mass movements in this environment. The study has been carried out on two scales, local and regional.

In studies of this type, attention may be focussed on the processes per se, that is, their mechanisms; on the environmental controls conditioning the activity of the processes; or upon the morphological response to the processes on the slopes (cf. Caine 1971). In the present context, the aim has not been to go into the geotechnical aspect of slope stability, or detailed snow physics with respect to avalanches, although some measurements and discussions relate to these approaches. The main problem has been mass movements from a geographic and geomorphological standpoint and the study deals primarily with the environmental controls and process responses (morphological results) of particular mass movements. The two types of slope processes particularly studied are examples of sporadic events, often of large

magnitude and referred to as 'catastrophic' processes (Starkel 1976). They recur only with many years intervals normally and therefore demand other methods than direct observations to be appropriately studied.

The objectives of the study can be stated in four points:

1. Describe the distribution of landforms caused by the investigated mass movements within northern Swedish Lapland and study relationships to environmental factors.

2. Study type cases with three aims: a) describe the characteristic morphology of process-related features, b) assess overall influence on slope form, c) infer process characteristics from morphological evidence.

3. Collect field data on recent and former activity or frequency of the mass movements, and estimate the relative geomorphological significance of the processes. The latter aspect includes an estimation of average magnitude (amount of erosion) of the processes, and of their recurrence intervals during postglacial time. Possible variations in activity related to climatic variations is a further issue worthy of attention in this context. Thus, the objectives under this point should also be used to evaluate the observed present distribution of process-related landforms with respect to the problem: is the occurrence of mass movement landforms in the Swedish mountain area primarily a result of widespread slope instability during a short period after the deglaciation, or of continued activity throughout Holocene?

4. Study the occurrence of debris flows and slush avalanches with reference to their role as natural hazards.

Hence, the investigation includes the two necessary steps in an analysis of erosion/sedimentation systems, according to Rapp (1975): a) studies of combinations of process types and resulting landforms, b) studies of magnitude and frequency of mass wasting processes.

As some of the landforms and geomorphic processes dealt with here are fairly little known, both in Scandinavia and elsewhere, an important part of the work has been the description and documentation of their characteristics and distribution. The investigation thus embraces both a systematic and a regional aspect. For comparative purposes, parts of Norway were visited in the course of the study.

1.3 The role of mass movements in periglacial slope development

In spite of the considerable interest in the periglacial environment shown by geomorphologists during this century, knowledge of periglacial slope systems, regarding function and change through time, is still limited. French (1976), in a general survey of the periglacial environment, ascribes this to four reasons:

1. a lack of quantitative measurements and field observation,
2. the slow rate of slope development, and therefore poor documentation of slope changes,
3. the difficulty of taking into account past climatic (non-periglacial) inputs to landform evolution,
4. the general problem of linking slope form and process.

Comparatively few studies have dealt with the relative importance of different geomorphological processes in northerly periglacial areas. With respect to mass wasting, these studies include works by Rapp (1960a,b), Jahn (1960,1961,1967), Rudberg (1962, 1963, 1964), Everett (1967), Washburn (1967), Gray (1972), Strömquist (1973,1983), French (1974), Luckman (1977,1978), Åkerman (1980, 1984) and Söderman (1980).

In periglacial areas of high mountains in lower latitudes, several studies of relevance have been made in continental Europe, e.g. Jäckli (1957), Veyret (1959), Schweizer (1968), and Kotarba (1976). Similar work has also been carried out in Colorado Rocky Mountains of USA (e.g. Benedict 1970, Caine 1976).

In a recent survey of processes modifying the Fennoscandian landscape in postglacial time, Rudberg (1984) concludes that rapid mass movements 'give substantial results in mountain areas' and that slow mass movements there appear to be faster compared with temperate areas. Generally, mass movements have been held as of particularly great significance in periglacial areas (e.g. Tricart 1970, Bird 1974, Caine 1974, Embleton & King 1975, Carson 1976, French 1976). Among reasons given to support this assumption are:

a) the importance of frost action, causing heave and sorting of materials on slopes, whereby downslope movement is accomplished on thawing by the influence of gravity. The process of frost creep is therefore thought very effective.

b) the effects of snow melting and ground ice thawing under conditions of impeded drainage due to frozen ground at depth, causing high moisture in the surficial layers. In suitable materials, this facilitates the process of gelifluction or induces rapid mass movements.

c) the generally sparse vegetation cover, which only marginally contributes to the stability of surficial soil layers.

However, actual measurements of e.g. rockfall rates and cliff recession do not show exceptionally high values for periglacial areas (French 1976 p.148). On the other hand, movements in unconsolidated material seem to be more widespread and probably higher in arctic/alpine areas in comparison with temperate ones (Benedict 1970, Young 1972).

Mass wasting on periglacial slopes operates with a marked temporal and spatial variability, conditioned by climate and terrain characteristics. The distribution of landforms resulting from specific processes gives testimony of the importance of the spatial regulators on process activity. Of importance in this context are i.a. slope angle and material as well as the availability of water, the latter largely governed by the snow cover regime, which is of great significance in periglacial morphogenesis (Thorn 1978). Rapid mass movements are characterized by their irregular and often low frequency of occurrence, as well as marked localization to steep slopes. In the periglacial environment, a particular suite of mass movements are those formed by rapidly moving snow or ice (avalanches), capable of eroding and transporting rock debris (Luckman 1977).

The evolution of slopes in most periglacial areas is progressively modifying forms inherited from a glacial climate or possibly

even older subtropical climate (Büdel 1977). On rock walls, the main result of periglacial slope development up to the present is probably the enlargement of chute systems or nivational cirques, and a limited general rock wall retreat, with concomitant building of debris slopes and cones below the walls. In arctic areas, rock walls with a height of several hundred meters have been buried by debris slopes in postglacial time, according to Bird (1974, p.705). On slopes in glacial deposits, periglacial mass wasting gradually accomplishes sorting and slight downslope movement of surficial layers.

The characteristic evolution of slopes associated with periglacial conditions is probably incomplete in most present-day periglacial areas (Carson & Kirkby 1972 p.327, French 1976 p.162), due to the repeated interruptions of periglacial morphogenesis by glacial regimes. Unglaciaded continental tundra areas in N. America and Siberia should have the most advanced development of slopes under periglacial conditions, but have not been widely studied.

To assess the relative role of different mass movements in periglacial slope evolution, process rate studies on a catchment basis (e.g. Rapp 1960b) and detailed inventories of the distribution of the features on different scales (e.g. Rudberg 1962, Åkerman 1980, Kotarba 1985) are of particular interest. Magnitude-frequency studies of catastrophic processes are required as a basis for an evaluation of the significance of rare extreme events in comparison with slow, continuous processes. The importance of mass movements versus other processes, primarily running water, is a further problem facing the geomorphologist working in the periglacial environment.

2. Study area

2.1 Regional scale: the northern Lapland mountains

2.1.1 Location

The regional study area is situated in the western part of northern Swedish Lapland, between approximately 66°N - 69°N and 16°E - 21°E , cf. Map III. An area of about 30 000 km² has been surveyed concerning the investigated mass movements by air photo and map analysis. Field work has mainly been carried out in the region south of lake Torneträsk (Cf. chapter 3). For practical reasons the investigated area embraces the mountainous part of the province of Norrbotten which borders on Norway to the west and on the province of Västerbotten to the south. The eastern boundary of the area was chosen as a line separating the mountainous terrain usually above the tree line (Swedish 'kalfjäll') from the flatter forest areas, partly dictated by the availability of background information such as geomorphological maps.

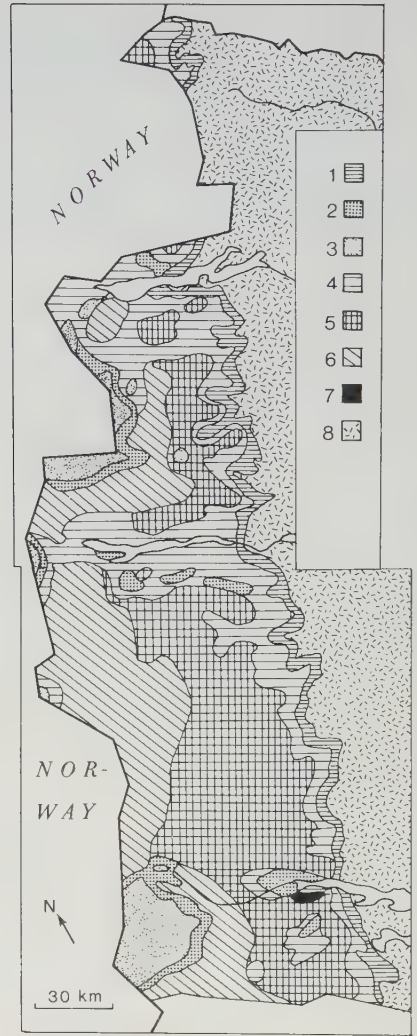
2.1.2 Geology and geomorphology

The area mainly belongs to the northern part of the Scandinavian mountain chain (the Scandinavian Caledonides). The easternmost parts of the area are situated on the crystalline basement rocks of the Baltic shield. Within the Norrbotten mountains at least three major tectonic units have been identified in the form of overlapping thrust nappes of westerly origin (Fig. 2.1). Morphologically this shows up as gentle slopes dipping westwards from east-facing scarps, representing the nappe fronts. Through erosion parts of the nappes have been locally removed west of their eastern boundaries, with the exposure of 'windows' of lower nappe units or the basement. The most extensive tectonic unit is the upper Allochthon, composed of several subunits partly of Caledonian igneous rocks. The subunits show varying grades of metamorphism, higher grade (the Seve Nappes) generally occurring below lower grade (the Köli Nappes). The lithology of the former is characterized by amphibolites, gneisses and schists, among which the amphibolites form distinct high relief areas in the terrain. The Köli Nappes consist of a volcano-sedimentary sequence of schists, greywackes, dolomites and other rocks, which often make up low relief areas to the west of the high amphibolite massifs (Kulling 1964, 1982, Gee & Zachrisson 1979).

The present-day relief is considered to originate from weathering and fluvial activity in the warm Tertiary and repeated glaciations in the colder Quaternary. Periglacial features of younger age show an altitudinal and latitudinal zonation of climatic origin, (Rudberg 1972, 1977). Different generations of erosion surfaces have been described from the southern part of Lapland (Västerbotten) by Rudberg (1954) and constitute ancient forms in the landscape. Alpine morphology with glacial trough valleys, cirques etc. dominates the higher massifs, in particular the Sarek and Kebnekaise areas, where a few hundred recent glaciers

Fig. 2.1. Generalised map of the structural geology of the study area, showing the main overlapping thrust units of the Caledonian orogeny. Redrawn from Gee and Zachrisson (1979). Key:

1. Parautochthon (Torneträsk group, sandstones, shales)
2. Lower allochthon (quartzites, shales a.o.)
3. Lower allochthon (Precambrian basement, granites, quartzites a.o.)
4. Middle allochthon (Akkajaure and Abisko complex, sandstones, mylonitized granites, quartz schists a.o.)
5. Upper allochthon (Seve nappes, amphibolites, schists a.o.)
6. Upper allochthon (Köli nappes, schists, quartzites, marble a.o.)
7. Upper allochthon (Särv nappe, sandstones and dolerites)
8. Autochthonous basement rocks (granites, gneisses a.o.)



exist (Östrem et al. 1973). Broadly speaking, the study area can be divided into a central zone of alpine geomorphology, and areas of lower rounded mountains west- and eastwards (Fig. 2.2). A set of geomorphological maps (scale 1:250 000) has been published covering the Swedish mountains (see Hoppe 1983). Especially concerning the medium-scale landforms of the study area, the reader is referred to these maps and descriptions, appearing in the SNV Pm-series (Swedish National Environment Protection Board).

2.1.3 Vegetation and soils

Two latitudinal vegetation zones occupy the major part of the area: the arcto-alpine zone north of Lake Torneträsk and the alpine zone to the south (NU 1983:2). Eastwards from the mountains commences the northern boreal zone of coniferous forest. The species composition of the zones is dominated by northern and

northeastern species, and by typical mountain species. The number of vascular plants in the region is around 300-400 (Op.cit.).

Five altitudinal belts can be distinguished. Following Rune (1965), they are:

1. the 'pre-alpine' boreal conifer forest belt to the east of the mountains.
2. the sub-alpine birchwood belt (*Betula pubescens* spp. *tortuosa*), which replaces the conifer belt at elevations above about 500-700 m a.s.l. in Swedish Lappland, forming the tree line.
3. the low-alpine belt of willow shrub and lower vegetation above the tree line.
4. the middle-alpine belt of mainly low alpine plants.
5. the high-alpine belt of mainly mosses and lichens.

This zonation follows two main gradients, one N-S and one E-W, due to broad climatic factors influencing the growth conditions of the vegetation. Within the belts there occur variations in these conditions principally caused by the type of soils and bedrock and climatic factors on the local scale due to topographic and aspectual effects. The belts are shifted towards lower altitudes in progressively more northerly and westerly areas. The tundra ecosystem characteristic of this part of northern Scandinavia has been the subject of several studies within the IBP investigations (Sonesson 1980a). The soils are of three main types: lithosols in the high mountains, podsoles and organogenic soils in major mountain valleys and lower forested areas. The mineral soils have their origin in silicious or base-rich parent rocks.

Detailed vegetation mapping (scale 1:100 000) based on interpretation of infrared air photos is available for the Swedish mountain areas, see e.g. SNV (1981).

2.1.4 Climate and hydrology

The general climatic characteristics of northern Swedish Lappland are briefly outlined below. Further data on the climate within local areas during the study period are given in sections 4.3 and 5.3.

The climatic conditions of northern Lappland are in a general way typical of mountainous regions, with a complex interaction of three geographical factors, latitude, altitude and topography, influencing the usually highly variable weather characteristics on a local scale (cf. Barry 1981 ch.2). The latitudinal factor (northerly position) is partly counteracted by the favourable influence of the Gulf Stream. The roughly north-south extension of the mountain chain along the Norwegian coast and the prevailing westerly wind regime give rise to a pronounced orographic effect and strong west-east gradient in cloudiness and precipitation, with higher values to the west. In the study area, a maritime influence is noticeable in the western parts, which from a climatic viewpoint constitute a transition zone between mari-

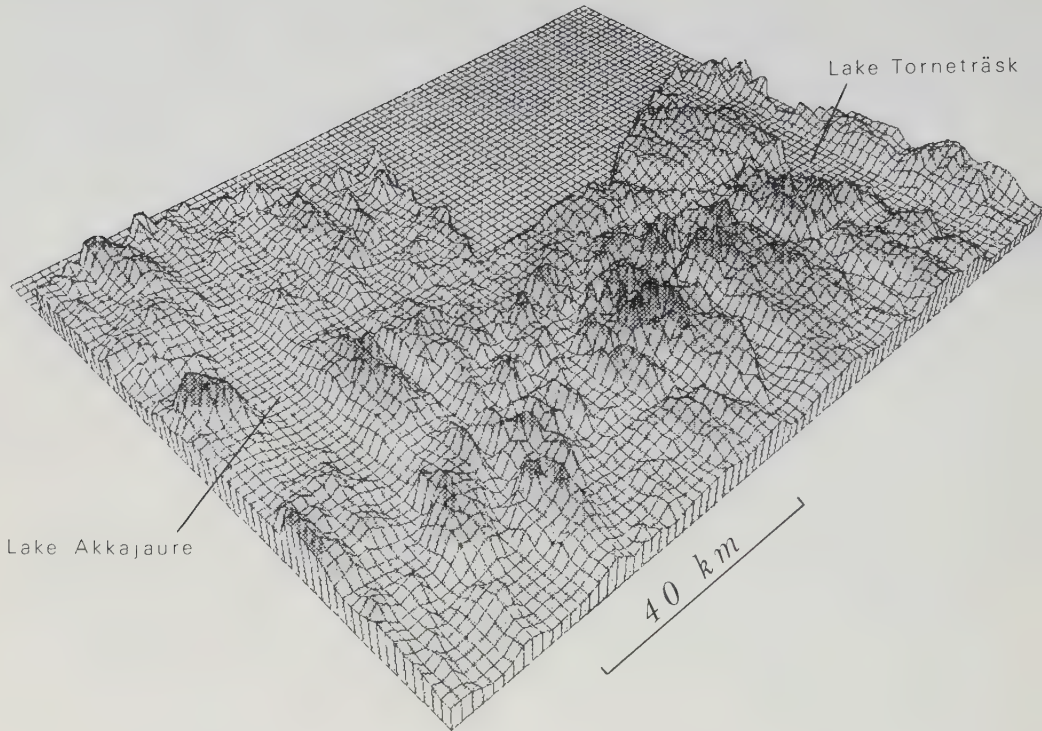


Fig. 2.2. Digital terrain model of the northern Lapland mountains, from the Stora Sjöfallet area in the south to Lake Torneträsk in the north. Dark areas are above 1200 m a.s.l. Norwegian territory shown flat.

time and more continental conditions eastwards (Wallén 1960). This is brought out by temperature and precipitation data for different stations in the area, shown in Table 2.1. The high precipitation amounts in the westerly parts are mainly connected with strong cyclonic activity. Eastwards some areas lie in rainshadow and have low cloudiness and precipitation, e.g. around Abisko.

In the Köppen climate classification, three climatic types cover this part of northern Scandinavia: Cfc (temperate rainy climate with cool short summer, along the Norwegian coast), Dfc (snowy-forest climate with cool summers, eastwards of the coastal zone at low altitudes) and finally ET (polar tundra climate with warmest month mean temperature under 10°C , making up the central high mountain region), cf. Strahler and Strahler (1978, plate D4). As can be seen from Table 2.1, mean temperatures of the warmest month are slightly above the criterion for ET climates for the weather stations of the study area, situated at low altitudes. This is a general problem with climatic data for mountainous regions, where measurements are normally made in valley locations, rather than on the mountains themselves. Only isolated studies are available for an appreciation of the climate of the high mountains proper (e.g. Hamberg 1901).

Table 2.1

PRECIPITATION Normal values 1951-80 in mm of monthly and annual precipitation for some stations in the study area. Source Eriksson 1982.

station	Lat N	Long E	Altitude m a. s. l.	Precipitation, uncorrected																Corr. ann. sums	% solid precip.
	J	F		M	A	M	J	J	A	S	O	N	D	Year							
Riksgränsen	6826	1808	508	80	64	56	51	47	75	93	99	112	119	72	72	940	1425	51			
Katterjåkk	6825	1810	515	69	55	48	44	40	65	80	85	96	102	62	61	807	1230	51			
Låktatjåkko	6822	1828	1220	220	170	150	130	80	80	100	110	120	220	180	190	1750	2910	80			
Abisko	6820	1850	388	25	19	19	12	14	28	52	47	28	29	25	24	322	460	40			
Torneträsk	6813	1943	393	27	21	22	19	27	41	75	66	39	32	30	29	428	520	42			
Nikkaluokta	6751	1902	470	32	25	25	33	32	43	81	73	50	35	39	35	503	685	50			
Kvikkjokk	6657	1745	337	37	30	26	29	36	54	85	76	60	43	44	47	568	705	43			

TEMPERATURE Monthly and annual temperatures for the normal period 1951-80 for stations in the study area. Source Eriksson 1982.

station	J	F	M	A	M	J	J	A	S	O	N	D	Year
Riksgränsen	-11.3	-11.3	-8.3	-4.1	1.2	7.1	10.5	9.1	4.6	-0.3	-6.0	-9.6	-1.5
Katterjåkk	-11.7	-11.7	-8.7	-4.6	1.0	7.0	10.5	9.0	4.4	-0.6	-6.3	-9.8	-1.7
Abisko	-11.7	-11.6	-7.9	-3.4	2.2	8.2	11.3	9.8	5.3	0.4	-5.4	-9.1	-0.9
Torneträsk	-12.3	-12.6	-7.9	-3.3	2.6	9.0	12.3	10.3	5.4	0.0	-6.4	-10.3	-1.0
Nikkaluokta	-16.0	-15.0	-9.5	-4.0	2.9	9.4	12.0	10.2	4.7	-1.7	-9.5	-13.9	-2.4
Kvikkjokk	-14.9	-13.5	-7.5	-1.7	4.4	10.5	12.9	11.2	5.6	-0.3	-7.7	-12.8	-1.1

Fig. 2.3 displays graphically mean monthly temperature and precipitation for Abisko. Mean temperatures are below freezing during six months of the year (November-April) and close to zero during October. July is the warmest month, and has also the maximum amount of precipitation. For Abisko and other continentally-influenced stations the precipitation is highest during the summer months. A different pattern is shown by west-located stations (Riksgränsen, Katterjåkk, Låktatjåkka), where precipitation maxima occur in autumn (October). These stations also have a smaller annual temperature amplitude reflecting their maritime character. Generally, yearly mean temperatures in the northern Lapland mountains fall within 0° to -4°C and mean precipitation within 500-4000 mm (NU 1983:2). About 50% of the precipitation falls as snow for the valley stations, considerably more (80%) for the single high mountain station of Table 2.1 (Låktatjåkka). The snow cover in the region lasts between 220-240 days of the year (Ångström 1974). Snow depths usually reach a maximum of between 0.75-1.5 m in March-April (Pershagen 1981). Variations between different years and stations are considerable, as shown by Fig. 2.4. Strong wind drifting causes uneven distribution of the snow cover above the tree limit with preferential accumulation on east-exposed slopes, which are favourable sites for nivation and local glaciation, as well as being particularly prone to avalanching.

Climatic variations during the Holocene are known in general outline and show a fluctuating pattern between warmer/drier and cooler/wetter periods, separated by about 200-300 years (Karlén 1982), cf. Fig. 4.38.

Stream flow is typical of periglacial regions, with a long largely inactive winter period and a short intense period of snowmelt runoff in spring. In the Torneträsk region, peak discharge usually occurs in June (Ekman 1957). Stream breakup is sometimes associated with slush avalanching. Occasionally,

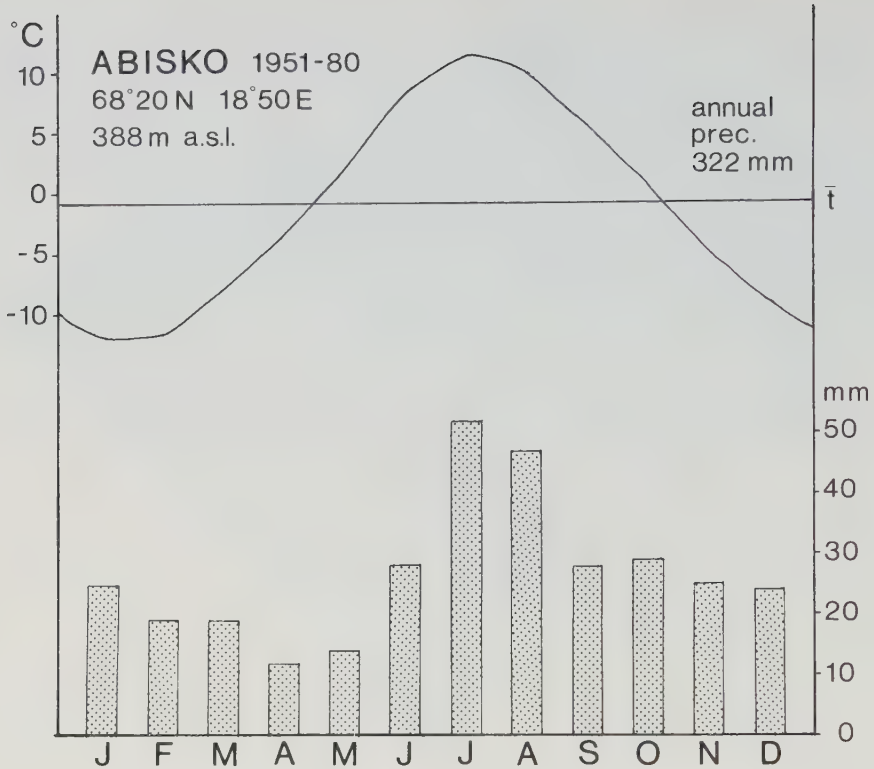


Fig. 2.3. Mean monthly temperature and precipitation at Abisko. After Eriksson 1982.

temporary runoff peaks caused by heavy rainstorms occur in the summer or autumn.

Slope wash is widespread during brief periods at melting in the spring before the ground is thawed. Downslope of long-lasting snow fields or glaciers sustained runoff may occur all through the summer season, with variations in intensity related to local weather conditions. The occurrence of frozen ground, partly as permafrost (Op. cit.) is important with respect to the type of runoff operating, promoting high pore water pressures in upper thawed soil layers and surficial flow on slopes. Later in the summer, when the ground is thawed to greater depth or completely, interflow becomes more important as a runoff process. The hydrological conditions are of significance in the context of erosion occurring on slopes. Rapid mass movements may be released after water inputs resulting from rapid snow melt or heavy rain showers.

2.1.5 Human influence

The impact of human activities on the environment is comparatively moderate in the Swedish Lappland mountains, cf. the geomorphological maps, e.g. Ulfstedt (1980). Reindeer breeding by the Lapplandish population has a long tradition in the area. In 1903, the railway Kiruna-Narvik was opened, facilitating access to the Torneträsk region. Recently, a highway along the same route was completed. A few roads into the mountain area exist also further

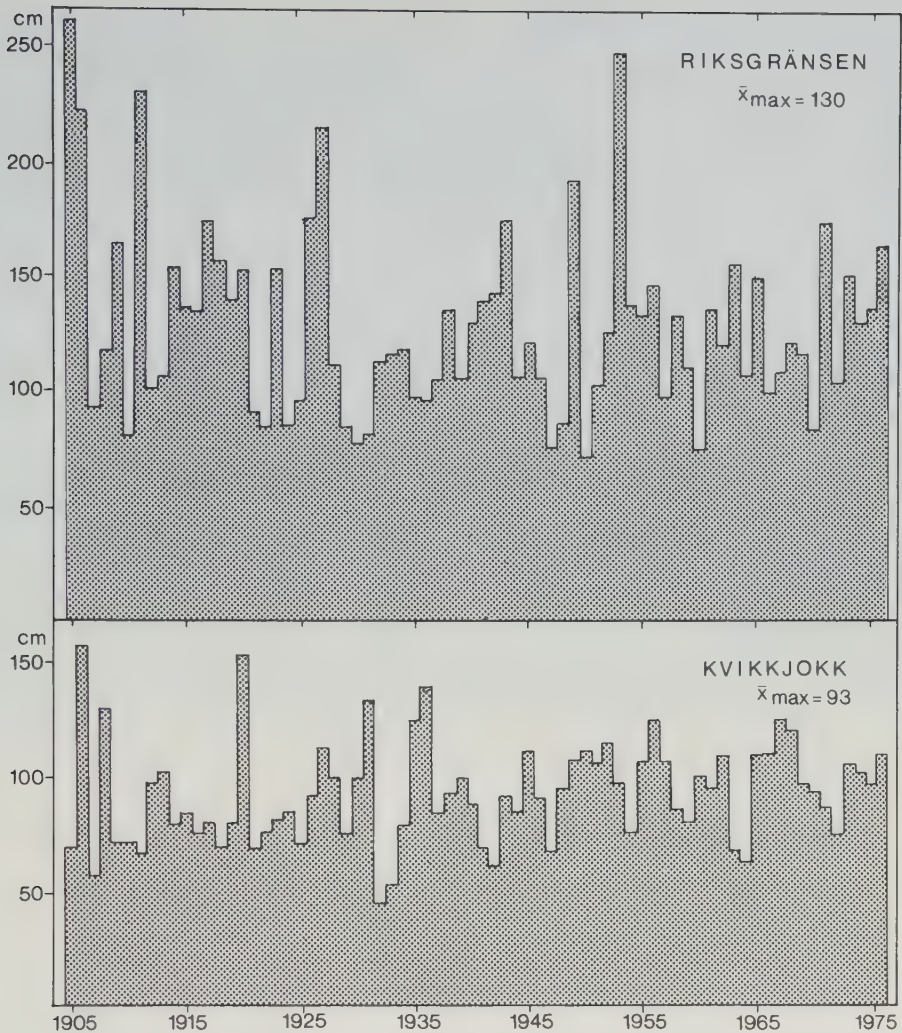


Fig. 2.4. Maximum snow depths at Riksgränsen and Kvikkjokk during the period 1905–1976. Riksgränsen exemplifies a snow-rich westerly station with maritime influence. Kvikkjokk has a more easterly location and continentally influenced climate. A great variability in maximum snow depth between different years is evident, particularly for Riksgränsen. After Pershagen 1981.

to the south. The impact on the environment following the construction of the Kiruna-Narvik road are dealt with by geographers in several studies (Gretener et al. 1981, Rehn et al. 1982). The recent trends in the tourism in the mountain area have also been investigated (Bäck & Hedlund 1982). A continuous monitoring of various aspects of the natural environment is furthermore provided by the activities of the scientific research stations in the area, at Tarfala (glaciology) and at Abisko (ecology, climate i.a.), cf. Schytt (1973), Sonesson (1980b). With respect to geomorphology, most of the area is to be regarded as natural and can be studied largely without the need to consider human influence.

2.2 Local scale: Nissunvagge

The Nissunvagge area was chosen for a more detailed study of mass movement impact on the local scale because of a recent event of geomorphic activity, which offered a favourable opportunity to study erosional effects and resulting landforms in a fresh state. Some parallel studies were undertaken in the Tarfala valley, which will be reported on in another publication.

2.2.1 Location

Nissunvagge is situated about 10 km SE of Abisko. It is a glacial trough valley cutting through the amphibolite massif known as the 'Abisko alps'. The general location is shown in Figs. 3.1 and 4.2.

2.2.2 Geology and geomorphology

The topography of the central part of the valley is depicted on Map I. The geology is dominated by amphibolitic rocks of the Seve nappe with gently NW-dipping planes, visible on the rock walls as alternating bands of more and less resistant bedrock outcrops (Fig. 2.5). The general features of the geology of the area are shown in Fig. 2.6, after Kulling (1964). Detailed geological mapping is still in progress in the Abisko mountains (M. Lindström and collaborators, Marburg). During the repeated Quaternary glaciations, the valley has been widened and deepened to the typical glacially-eroded U-profile (Fig. 2.7). The main continental ice flow during the latest glaciation had a general SE-NW direction across the area (cf. Melander 1977). Local cirque glaciers presently exist at Kukkesriepe and Kaskariepe tributary valleys.

Till is the dominating loose deposit in the valley bottom and on slopes, although in the latter case locally covered with postglacially-produced talus material (cf. Map II). Patterned ground and gelifluction lobes usually characterize the surficial aspect of the till cover. On the higher mountain slopes and ridges, extensive block fields occur. Glacifluvial deposits associated with the deglaciation occur in valley-bottom positions or in lateral more elevated sites, sometimes connected to drainage channels of ice-marginal origin. Deposits of organic material in fens occupy minor areas in terrain depressions, locally with sporadic permafrost in palsas.

2.2.3 Vegetation

Nissunvagge is a typical high-mountain valley, wholly situated above the tree line. The nearest birch forest is to be found along the Nissunjokk canyon to the north reaching an altitude of about 780 m on the sun-exposed side. The only higher vegetation in Nissunvagge is a belt of Salix extending from the canyon a few km southwards mainly on the eastern valley side. The major part of the valley is dominated by either meadow type of vegetation, characterized by grasses and herbs, or heath vegetation with low shrubs of Betula nana and Empetrum hermaphroditum. On wind-exposed hills or glacifluvial terraces, there exists a dry heath distinguished by patchy vegetation of lichens and e.g. Loiseleuria.



Fig. 2.5. Westfacing slope of Mt. Nissuntjärro, Nissunvagge. The morphology has a strong structural background, being due to alternating bands of bedrock (amphibolite) with variable resistance. The structural weaknesses have guided erosion, causing faster retreat in some layers than for the wall as a whole. Strong layers are therefore undercut, and preserve a steep slope angle during parallel retreat. A characteristic feature of the rockwall is the chute sculpture, caused by selective erosion along joints of the bedrock, and of great importance in channelling of debris transported by mass movements. A large number of debris flows from 1979 can be seen on the talus slope beneath the rock wall. Air photo N.Å. Andersson, Abisko, 31.7.1980.

A description of the composition of vegetation characterizing steep slopes with respect to aspect will be given in section 4.3. For further information on the distribution of the vegetation of Nissunvagge and surroundings, see the vegetation map of the Swedish mountains No.2 Abisko, (SNV 1981).

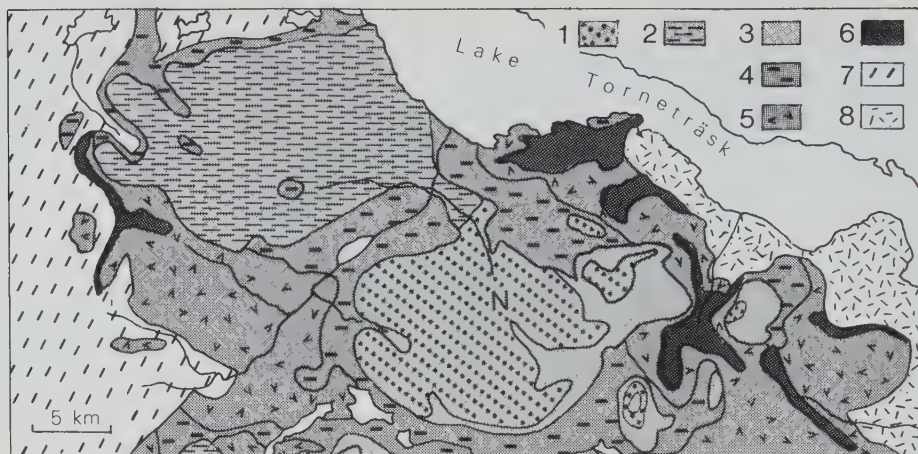


Fig. 2.6. Geology of the Abisko mountains south of Lake Torneträsk, including the Nissunvagge area (N). Key: 1. amphibolite 2. mica-, quartzose-, graphite-schist, marble 3. gneiss 4. quartzitic and phyllitic rocks ('hard schists') 5. igneous rocks, mainly granite and syenite 6. quartzite, phyllitic slate 7. Archaean basement in windows (granite, mica schist a.o.) 8. Archaean basement to the east of the Caledonides (granites a.o.). After Kulling (1964).



Fig. 2.7. Nissunvagge seen from the south. The glacial morphology of the valley is evident in the u-shaped profile, and the hanging cirque to the left. The level mountain top plateaus and peaks, often with marked summit accordance in height, presumably represent remnants of an older erosion surface of preglacial age. Peak in the background is Mt. Pallentjåkka, mountain to the right Mt. Nissuntjärro (cf. Map I). Air photo N.Å.Andersson 31.7.1980.

2.2.4 Climate

The common occurrence of low vegetation adapted to strong winds, and the existence of permafrost, sporadic at present as evidenced by palsas in the valley bottom, but previously more extensive as witnessed by fossil ice wedge polygons, indicate a cold, continentally-influenced climate. The nearest weather station is at Abisko by lake Torneträsk more than 10 km to the northwest and at a much lower altitude. No data were available for Nissunvagge prior to this study. Some limited measurements were therefore carried out in connection with the field work in the valley. The measurement site was located at 950 m altitude on a W-exposed slope in the central part of Nissunvagge (cf. Fig. 6.2). Air temperatures for Nissunvagge (950 m a.s.l.) and Abisko (370 m a.s.l.) during the snowfree season of 1982-83 are shown in Fig. 2.8, based on thermohygrograph-measurements in standard meteorological screens. These data indicate lower temperatures of generally about 3-4°C in Nissunvagge during spring and early summer, closely corresponding to the expected temperature difference under normally-prevailing lapse rates of 0.5-0.7 °C/100 m (Ångström 1974). During the later part of the summer and during autumn (September- October), the temperature difference tended to be somewhat larger, possibly related to a warm temperature lag in Abisko, not felt in Nissunvagge, caused by Lake Torneträsk. Occasional extreme contrasts exceeding 10° are furthermore noticeable in Fig. 2.8, as well as brief periods of warmer temperature in the mountain locality compared with Abisko. An estimate of mean annual air temperature for Nissunvagge at 1000 m altitude, based on Fig. 3 in Eriksson (1982) is about -3°C, and for the summit altitude of Mt. Nissuntjärro (1738 m) about -5°C.

The number of freeze-thaw days during the measurement periods (Table 2.2) was much higher in Nissunvagge than in Abisko, except for the late autumn 1983 (15 Sept-14 Oct), when temperatures dropped below zero for an extended period in the mountain valley, whereas Abisko had numerous freeze-thaw changes. Cycle frequency in Nissunvagge had peaks of similar magnitude in spring and autumn. Cycles in July were absent for the measurement site. Data on precipitation for a limited series of days are given in Table 2.3, compared with the corresponding data for Abisko. Rainfall in Nissunvagge was measured using a SMHI-gauge at ground level. No values are currently available for annual precipitation in the valley. The available data suggest a substantially higher amount of rainfall during the summer period compared to Abisko, which has exceptionally low values due to a rainshadow effect.

Snow precipitation is probably fairly low and strongly redistributed by wind drifting. Visits to the valley in the winter strengthen this assumption. The snow cover is discontinuous with local bare-blown spots on ridges and slopes, and often less than 0.5 m deep. However, locally in lee positions to westerly winds below mountain crests, huge snow drifts accumulate yearly, some large enough to survive the summers (cf. Map I). Evidence for strong winds in the winter is obtained from surface features of the snow cover. Wind indicators show a strong predominance of directions from NNW and SSE, thus along the valley alignment.

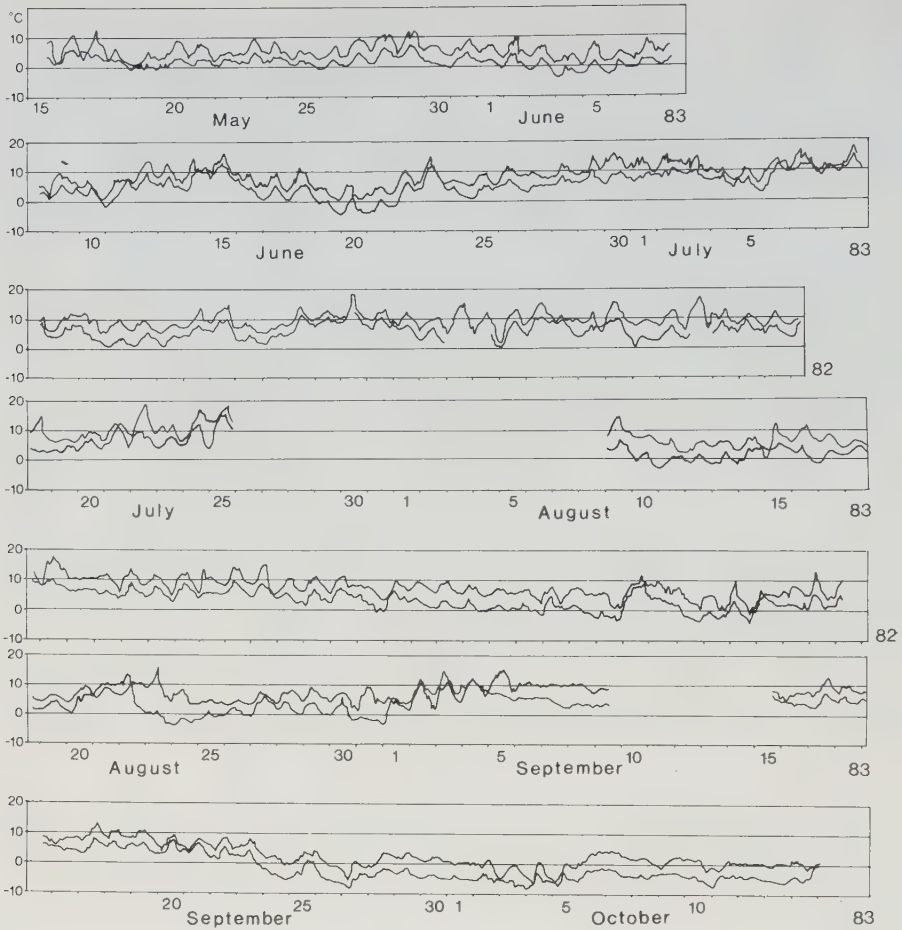


Fig. 2.8. Temperature data for Nissunvagge at 950 m altitude (lower curves) and Abisko 370 m altitude during the measurement periods in 1982 and 1983. The record for Nissunvagge is partly incomplete. Data from meteorological screens with THGs.

Table 2.2. Number of freeze-thaw days for Nissunvagge and Abisko during measurement periods in 1982 and 1983

period	freeze-thaw days	
	Nissunvagge	Abisko
August 1982	1	0
1-18 September 1982	6	1
15-31 May 1983	4	0
June 1983	11	0
8 August-8 September 1983	11	0
15 September-14 October 1983	3	13

Table 2.3. Precipitation data for Nissunvagge (950 m altitude) compared with Abisko (370 m altitude).

Daily recordings 7 a.m.

	mm rainfall	
	Nissunvagge	Abisko
July 1982		
11	2.0	0.2
15	-	-
16	-	-
17	1.5	5.1
18	28.2	13.8
19	-	0.5
20	2.0	0.2
21	-	-
22	-	-
23	2.0	-
August 1982		
18	-	-
19	-	-
20	2.7	0.0
21	20.0	0.5

Periodic recordings

	Nissunvagge ¹	Abisko
1982		
21.7-18.8	68.3	67.9
19.8-21.8	22.7	0.5
22.8-18.9	42	26.1
total period 21.7-18.9	133	94.5
percentage difference	140 %	
1983		
8.6-9.8	199	166.0
10.8-6.9	59.6	71.1
7.9-15.9	73	13.4
total period 8.6-15.9	331.6	203.7
percentage difference	163 %	

^{1/} rainfall gauge (SMHI) left unattended for long periods, emptied sporadically. No correction made for evaporative loss.

3. Methods

3.1 Field work

Data collection and description of the mass movements have mainly been carried out in the field. The extent of field visits and the location of major sites mentioned in the text are shown in Fig. 3.1. The field methods used are listed in Table 3.1. Most of these need no further description here. Some information is provided in connection with the presentation of the results in later chapters. However, two groups of methods used in snow studies and for vegetation dating will be dealt with separately below.

Table 3.1 Overview of field methods used in the study

<u>objective</u>	<u>method</u>
landform description	size and orientation measurements on surface materials, terrain profiling and measuring using clinometer, tape, compass plane table mapping sampling of material, study of stratigraphy
process observation	repeated measurements of lines of inserted pegs or painted stones to check slope movements installation of paint markings in depositional zones of mass movements for future checks of debris transport and accumulation monitoring of climatic data during research periods, using rain gauge, thermohygrograph, ground thermistors and thermometers snow pit studies of stratification, density and Ram penetration at avalanche sites experiments on snow strength in relation to free water content
dating of form or process	measurements on Rhizocarpon lichens and collection of material for dating by dendrochronology or C14-method



Fig. 3.1. Main field area of the study. Field traverses (dashed lines) and major localities (x) marked.

3.2 Numerical analyses

Mainly non-parametrical statistics have been used in some analyses of quantitative data. Most of the calculations or tests were made by the MINITAB-package (Ryan et al. 1981) available at the Lund University Computing Centre.

3.3 Cartographic and remote sensing applications

Methods of cartography and aerial photography have been essential as a complement to the field methods. Application of these techniques has a long tradition at the Department of Physical Geography in Lund, leading up to the present strong involvement

in environmental monitoring methods based on remotely-sensed data, and establishment of a well equipped Remote Sensing Laboratory. Four objectives were met by the use of these methods:

- A. to prepare base maps photogrammetrically.
- B. to survey local and regional distribution of the studied objects.
- C. to measure photogrammetrically slope gradients in air photos.
- D. to illustrate terrain relief and morphology in three-dimensional diagrams (digital terrain models).

A. Photogrammetric mapping.

One objective of the study was the cartographic documentation of the rapid mass movements that occurred in 1979 near Abisko. Affected slopes had been surveyed from the air on oblique photos of variable scales (N-Å. Andersson, Abisko Scientific Research Station). A special map was prepared photogrammetrically for the area most severely affected by the mass movements and used to chart their distribution from the oblique air photos. The map was made to a scale of 1:10 000, with contour interval of 10 m, using vertical air photos of scale 1:30 000 and a Wild Aviograph B8S. The map, printed to a scale of approximately 1:16 700, does not claim absolute accuracy in elevation measures, since an adequate number of control points was not available within the area. To obtain horizontation of the stereo model, mountain peak heights and lake surface levels had to be used. However, the relative height differences and the detail form of slopes are well represented. The map furthermore served as a base map to a survey of slope forms and processes in central Nissunvagge, based on air photo interpretation and field checking. A photogrammetric mapping was also made of an area between Björkliden and Låktatjåkka in the western Abisko region to provide a detailed contour map of studied field sites.

B. Air photo interpretation

Air photos (black-and-white 1:30 000, IR-col. 1:60 000) have been the main source of information regarding the distribution of slush avalanches in northern Lapland. An initial study was made concerning the possible use of Landsat MSS imagery in this respect (Nyberg (1979,1980)). However, the spatial resolution (about 30-80 m) and limited quality and information content due to the prevailing degree of cloudiness and extent of snow cover made the imagery only partly useful. A more time-consuming survey had to be carried out utilizing vertical air photos with stereoscopic overlap. Several hundred photos covering the whole study area were interpreted regarding the occurrence of avalanche sites along streams. The features of interest were the paths with signs of marked geomorphic activity detectable in the air photos. Sites were plotted on topographic maps to the scale 1:100 000 and subsequently transferred to maps of smaller scale before presentation. The photo interpretation was mainly performed during a five-day visit to the Swedish Natural Land Survey (LMV) in Gävle.

Concerning the accuracy of the survey, a quantitative measure has not been possible to provide. Due to the size of the surveyed area and lack of roads, field checks had to be made on foot. It was therefore impossible to follow any systematic or random scheme for ground control, because of time shortage. However, a sort of 'path-sampling' was made by checking mapped sites and

distances in between during field traverses (cf. Fig.3.1). The regional distribution of debris flows was analyzed on geomorphological maps covering the study area (cf. section 4.5.1).

C. Photogrammetric measurements.

Measurements of slope gradients were made photogrammetrically using a procedure described by Mekel et al (1964). The method, which is based on distance and parallax measurements in vertical air photo stereo pairs, has an accuracy comparable to that achieved by simple field measurement using Abney levels etc., or about 1-2 degrees. Data were measured for slush avalanche paths along streams and for stream courses not affected by slush avalanches. The data base consisted of the air photos covering northern Lappland available at the Department in Lund, from which a large sample of avalanche paths and a smaller sample of unaffected stream courses were measured. In total, 95 sites were included. Although there is a slightly higher coverage of air photos for the main field area (Fig. 3.1) than for other parts of Lappland, the results obtained are thought valid also more generally, as the variability in relief and bedrock in northern Lappland was well represented in the covered area.

D. Digital terrain models

The availability in recent years of soft-ware packages designed to produce computer-generated 3D digital terrain models (DTM) enables the geomorphologist to visualize landforms or landscape relief in a convenient way, as a complement to contour maps. The 3D-plots included in this study were made by digitizing contour information from official maps and the large-scale topographic map specially produced for the investigation, using a HI-STATE precision coordinate digitizer at the Remote Sensing Laboratory of the Department. The irregularly distributed sampled heights were then interpolated to a regular matrix of height data (DTM) by use of an interpolation routine in the UNIRAS GEOPAK package (UNIRAS 1982a). The 3D-plots were finally drawn by a RASPAK-routine (UNIRAS 1982b) on an IDS-printer. The routines were executed at the Univac 1100 computer at Lund University Computing Centre and could be started as batch jobs via the DIAPAC-package developed at the Remote Sensing Laboratory.

3.4 Snow studies

Snowpack analyses covering a limited number of parameters were made, using conventional methodology (see e.g. Perla & Martinelli 1975, p.57, 217). The Ram penetrometer used for snow strength measurements is a graded metal tube that can be driven into the snow vertically by means of a hammer dropped down a guiderod. Knowing the mass of the hammer (H) and tube (T) and the fall height of the hammer (f), a relative strength index (Ram number, R) for the snow in different strata can be calculated by the formula

$$R = T + H + nfH/p$$

where n= the number of blows of the hammer
and p= penetration of the hammer.

The liquid water content of snow was estimated by the hand test

(cf. Perla & Martinelli 1975 p.218). In view of its presumed importance with respect to the occurrence of slush avalanches, it was also decided to carry out some quantitative measurements and experiments, despite the well-known difficulties (cf. Colbeck 1978). A comparatively simple and inexpensive method was used in the experiments, being a modified form of Bader's solution method for determination of liquid water content of snow (Morris 1981). It is based on measurement of the freezing point depression of a NaCl solution caused by its dilution by the liquid water of a snow sample. The measurements were made using a 0.7 l dewar and a Tastoherm digital thermometer, with snow samples of 500 ml added to a pre-prepared brine of 100 ml NaCl (concentration 10 % by weight). For a description of the calculations involved in the determination of the liquid water content see Morris (1981).

3.5 Dating techniques

3.5.1 Lichenometry

Lichenometry as a dating method has been used since about 1950 (Beschel e.g. 1973), and has found many applications by geomorphologists (see review by Webber & Andrews 1973). Although the technique has been questioned as an absolute dating technique (Jochimsen 1966, Worsley 1981a), it has given consistent results with information from other sources (e.g. Karlén 1982). In recent years, attempts have been made to standardize procedures and give the method a firmer basis (Lock et al. 1979, Innes 1982, 1984).

In northern Scandinavia, lichenometric studies have been made by i.a. Bergström (1954), Stork (1963) and Karlén (1973 and later). They used the technique to study glacier recession during the Holocene. The main purpose of the application of lichenometry in the present study was to obtain information on the relative ages of different debris flows, or the number of older events in postglacial time. Preliminary results were reported on by Rapp & Nyberg (1981a,b). Comprehensive overviews of the basic principles, problems and applications of lichenometry are available elsewhere (Beschel 1973, Webber & Andrews 1973, Lock et al. 1979), and these issues will therefore only be briefly touched upon here.

A. Techniques of measurement and sampling

Discussions on sampling techniques in lichenometry are found in Lock et al. (1979) and in Innes (1984, 1985a). When the present study was initiated, it was decided to adopt the methods used by Karlén (1973) and tentatively use his growth rate curve. Possible short-comings of this type of aggregate growth-curve for Rhizocarpon lichens have later been pointed out by Innes (1982). Thus, despite resulting backache, the whole population of lichens (more or less) on the substrates to be dated was searched for the largest individuals. The use of small sampling plots was seen as unrealistic considering the limited areas of the substrates in this study (cf. Gordon & Sharp 1983). Irregular, highly elongated and coalescent thalli were excluded from the measurements and only roughly circular specimens were measured with a slide rule or steel tape to the nearest mm (longest axis). Using the recently suggested nomenclature for Rhizocarpon lichens (Innes 1985c), the measurements refer to the subgenus Rhizocarpon and include section Alpicola and section Rhizocarpon.

To account for a possible overestimation of the age of the dated mass movements by including in the measurements anomalously large lichens, surviving the mass movements, at least 20 of the largest lichens found were recorded for each site. The largest thallus was only accepted as representative of the age of the deposit if it did not differ from the second largest by more than 10 mm in size. Otherwise it was regarded as inherited from an older substrate. Furthermore, not only the largest lichen was used to represent a substrate, but also the mean of the five largest, as proposed by Innes (1984). The probability of including inherited lichens in the measurements in the study area was assessed from inventories of fresh debris flows of 1979 (cf. Rapp & Nyberg 1981b). About 3 % of a few hundred blocks inspected had surviving lichens.

B. Factors causing dating inaccuracies

The possible error of including a lichen older than the mass movement event to be dated has been mentioned above. Other factors that may influence the accuracy of the lichenometric dating are related to the colonization time lag of lichens and variations in their growth rate. Colonization by Rhizocarpon lichens on fresh debris flow deposits is probably rapid, as the deposits generally occur within areas where already established lichens are common. Time lags seem to lie between 5-15 years, until visible lichens emerge (cf. Karlén 1975). Compared with the ages obtained for different old debris flows in this study, the uncertainty introduced by unknown colonization time lag is not very important.

Variations in lichen growth rate, which at least partly may invalidate growth curves constructed for lichens on the assumption of continuous growth, may be due to several reasons, see e.g. Beschel (1973), TenBrink (1973), Beschel & Weidick (1973). In northern Sweden, Karlén (1975) found no marked variations in growth rate of Rhizocarpon geographicum and Rhizocarpon alpicola within the mountain region, despite climatic differences within the area and altitudinal zonation. Measurements of lichen sizes on different rock substrates of similar age showed no systematic correlation of growth rate with substrate (Op. cit.). However, Innes (1985a) found variations in growth rate related to bedrock type in the Svartisen area. Hence, the growth curve for Rhizocarpon lichens constructed by Karlén (1973) may be somewhat crude, as data from other areas, e.g. southern Norway, also indicate that the growth rate of the involved species may differ following the initial rapid growth period (Innes 1982). Most of the lichen size measurements in the present study were made on a substrate of amphibolitic rocks. Two locations differed with a substrate of schists.

The growth of lichens may be restricted by available clast surface area (Gordon & Sharp 1983). Datings of debris flows with predominantly small clasts may hence give too low ages. Prefixed sampling procedures are liable to neglect variations in clast size and produce unrepresentative results (Op. cit.). The method of locating the single largest lichen by extensive search over whole units is not as strongly affected by the problem of substrate variability.

Debris flows may be influenced by snow avalanches periodically, and this could delay lichen growth over the deposits by overtur-

ning blocks intermittently. Flows with visible avalanche influence were avoided in the measurements.

3.5.2 Dendrochronology

A second approach that was tried for estimation of debris flow frequency was tree-ring dating of mountain birch growing along debris flow tracks. For a short general review of the technique see Worsley (1981b). Its application specifically to geomorphic problems has been treated by Alestalo (1971).

The impact of mass movements on tree growth is manifold (e.g. Shroder 1980 p.165). Concerning avalanches and debris flows influence ranges from complete felling of trees to minor damage by corrosion or burial of stemwood. A number of tree growth responses occur following the mass movement events, and can be used for dating purposes, e.g. sprouting or colonization in avalanche tracks with periodic activity. In the Abisko area, this technique was used by Larsson (1974) to estimate snow avalanche frequency on Mt. Njulla. Similar applications are numerous from other mountain areas e.g. Butler (1979).

The measurements made were designed to provide some information on the length of the period of stability and undisturbed tree growth preceding the recent slope failures treated in the study. For this purpose, cores were extracted at 1.5 m height from a number of the oldest trees felled or damaged by the recent mass movements, using an increment borer. The cores were subsequently examined in microscope. The error in age determination due to ambiguous growth rings is estimated at about 2-3 years in 50 or about 5 %.

3.5.3 C-14 dating.

Datings of organic material buried by mass movement processes were made by the Radiocarbon Dating Laboratory at the Department of Quaternary Geology (Håkansson 1982, 1984). Due to the possible presence of calcite and graphite, two of the datings were made on the NaOH-soluble fraction of the samples.

4. Debris flows

4.1 Introduction

Debris flows are rapid mass movements occurring on slopes and along stream channels usually in mountainous terrain. A comprehensive review of research on debris flows concerning physical characteristics, types of deposits and potential hazard is given by Costa (1984). Much of the work on debris flows is of a geo-technical nature (e.g. Johnson 1970, Takahashi 1981) or centered on their role as natural hazards (e.g. Hollingsworth & Kovacs 1981). In Sweden, debris flows and other rapid mass movements do not have the same impact on populated regions, with danger to people and property, as in many other countries. Thus, the impetus to study mass movements of the rapid type for this reason has not been very strong. In Norway, for example, several applied investigations have been made on mass movements including debris flows, as for instance that by Dahl et al (1981).

Zenzén (1926) was one of the first to describe a case of rapid mass movements (possibly debris flows) in Swedish Lapland. Rudberg (1950) made another pioneer study, describing slides or debris flows released by thunderstorm rain in 1947 near Lake Ajaure. A more systematic study of mass movements started in the 1950's with the work by Rapp (1959 and later). He described the geomorphological effects on slopes in Kärkevagge of a heavy rainfall on Oct 6, 1959, releasing numerous debris flows (referred to as mudflows). In a quantitative analysis of geomorphic processes operating on the slopes of Kärkevagge, mudflows and earth-slides ranked second in importance, after solute transport by running water, and thus were the most significant mass movement in the development of slopes in this locality (Rapp 1960b, p.186). However, the possibility that the measurement period may have coincided with an extreme event, causing values higher than normal for these processes is recognized. Follow-up studies in Kärkevagge (Nyberg & Brydsten, in Strömquist 1977) have shown continued occurrence of debris flows, although not as numerous as occurred in 1959. Two other reported cases of debris flows in northern Lapland, at Tarfala in 1972 and at Nissunvagge in 1979 (Rapp 1974, Rapp & Nyberg 1981a) further stress the significance of these mass movements and suggest a spatial pattern of occurrence on different scales, related to the type of initiating rainfall, from isolated cases to widespread release of numerous debris flows in whole regions (cf. Rapp & Nyberg 1981a).

Important observations have been made with respect to local topographic and general material influences in the release of debris flows in this environment (e.g. Rapp 1960b, 1974, Sulebak 1969, Strömquist 1976, Dahl et al 1981). The general distribution of debris flows in the Swedish mountains has been documented in the geomorphological maps issued recently (e.g. Melander 1977). However, the research is still in an initial stage concerning local and regional patterns of occurrence, the morphological variability of debris flows, the triggering amounts of rainfall, and possible changes in debris flow frequency during the postglacial period. It follows that conclusions on the regional geomorp-

hic significance of debris flows require access to more data than are presently available, both regarding recent cases of debris flows, and especially for older flows. It is the aim of the present study to add some new information on these issues.

4.2 General characteristics of debris flows

4.2.1 Definition and occurrence

In classifications of rapid mass movements, a distinction is usually made between movements taking place as slides, comprising a coherent unit of soil or rock moving along a slip plane without internal shearing, and flows, involving internal disintegration of the material (e.g. Sharpe 1938, Varnes 1958). A third principal type of rapid mass movement, falls, can also be distinguished (Rapp 1960b). These different modes of mass movement are related to material properties, especially water content, with higher moisture in flowing debris.

The amount of water in the material of debris flows can vary widely and the process grades into muddy streamflow (mud or flash floods) with successive increase in water content. Debris flow and mud flood may occur in close association at a site (e.g. Pierson 1980, Costa 1984). Many definitions of debris flows have been given and other terms such as mudflow and debris avalanche have been used for essentially the same or at least very similar mass movements. Some authors include them together with quick clay slides in a group called flowslides (e.g. Statham 1977).

Varnes (1958) gives a general definition of debris flows as being a form of rapid mass movement of a body of granular solids, water and air. Several types of flows known under different terms may be encompassed in this definition:

a. mudflows, consisting predominantly of fine particles from sand to clay. The term has been used also for coarser flows (e.g. Sharpe 1938, Rapp 1960b) and for considerably slower types of flow movements, cf. Statham (1977, p.90).

b. coarse debris flows, incl. debris avalanches, consisting largely of particles from gravel to boulders

c. till flows or sediment flows, consisting of supraglacial sediments of varying sizes

d. lahars, consisting of recent volcanic deposits

The present study is concerned with type b, the coarse debris flows in an arctic/alpine area. This type of flow can be further defined as a rapid mass movement of blocky, mixed debris of rock and soil by flow of wet, lobate mass (Rapp & Nyberg 1981b).

The occurrence of debris flows is not simply related to climate (water regime), but also to vegetation cover and available relief. The most common factor for debris flow occurrence is mountainous terrain conditions or at least locally steep slopes, although small flows ('micro-mudflows') may occur on slopes as gentle as 8-10° in arctic regions (Everett 1967). Types c and d are restricted to regions with glaciers and active volcanoes, respectively. Types a and b seem to occur mainly in semi-arid and alpine environments of mountainous character (e.g. Blackwelder

1928, Fryxell & Horberg 1947), although they are known also from other climatic zones, e.g. Ireland (Prior et al. 1968). Recently, debris flows have been identified on locally steep slopes at Söderåsen, southern Sweden (Rapp 1984).

4.2.2 Flowage of coarse debris

Our understanding is restricted with respect to the flow mechanism of wet, coarse debris (soil with high percentage of stones and blocks), which may flow in the manner of wet concrete or lava, often rapidly and on quite gentle slopes, and be able to carry huge boulders for considerable distances (e.g. Sharp & Nobles 1953). The debris load for these flows may be as high as 90% (Costa 1984) and they are thus particularly interesting from a geomorphic viewpoint. The mechanism of debris flow is poorly known partly due to the difficulty of observing the process at work. Small flows can be studied on the frontal parts of debris-covered glaciers (Lawson 1982), but are not comparable to flows on land surfaces in terms of transporting capacity. Their importance is mainly manifest in stratigraphic context, and not of great significance with respect to landforms and slope denudation.

Only recently, sophisticated technical equipment has made possible direct observation and recording of large debris flows in motion, and some results of these field observations in Japan are now available (Suwa & Okuda 1980).

Debris flows superficially resemble some rapid lava flows in their mode of flowage (Johnson 1970). Similar flowage has also been observed in debris consisting of hail (Metcalf 1975). There are only a few direct observations reported, but these (e.g. Johnson 1970, Sharp & Nobles 1953, Curry 1966, Wasson 1978, Pierson 1980), as well as indirect evidence from the resulting landforms and laboratory experiments (Johnson 1970), allow some general features characteristic of flowage of coarse debris to be stated:

A. Movement tends to occur in a series of pulsations or surges with variable time intervals in-between (e.g. Wasson 1978). These surges are related to temporary damming of the flow track by debris and subsequent breaching of the dam as further material presses on from upslope.

B. After a pulse of debris there usually follows a more fluid turbulent flow phase with lower concentration of large clasts (Johnson 1970, Pierson 1980).

C. In major, viscous flows, large boulders are carried on the surface of the flow, primarily at the front, 'floating' in a matrix of finer material (e.g. Takahashi 1981).

D. The surface of the flow is smooth in gross outline, indicating laminar flow rather than turbulent (Johnson 1970). However, turbulence has been noted in-between laminar flow-regimes for debris flows in New Zealand (Pierson 1980).

E. The central part of the flow can move at constant velocity, without internal shearing, as a rigid plug (Johnson 1970, Wasson 1978). Thus some clasts may be transported with little or no abrasion affecting them, and e.g. have preserved lichen cover after deposition.

F. Movement is usually maintained on slopes steeper than about 10° if no blockages occur. On more gentle slopes, the flows tend to be drained of water and regain frictional strength, and thus come to a halt (Statham 1977).

Early research on debris flows tried to apply hydrodynamic models used for water flow to explain debris flowage (e.g. Sharp & Nobles 1953). Subsequently it was realized that rheological models would be more in sympathy with both field observations and laboratory experiments. Thus, flowage of coarse debris is described as plastico-viscous by Johnson (1970). Alternatively, Takahashi (1981) suggested that a dilatant-fluid model can be used to describe debris flows, using the concept of dispersive pressure between flowing particles to explain transporting capacity and concentration of boulders at the surface and front of a debris flow. A number of physical factors relating to the material of debris flows have been considered as important in explaining the flow properties of the debris:

- 1) the presence of a clay fraction. Mixed with water, clay particles provide a cohesive slurry that supports and reduces effective stresses between other particles (Rodine & Johnson 1976).

- 2) poor sorting of the debris. This leads to low interlocking of clasts, which decreases strength and increases mobility of the debris (Op.cit.).

- 3) high density of the debris. This favours buoyancy effects, due to weight of displaced debris, for boulders, which are thereby more easily transported (op.cit.).

- 4) high pore water pressures. The weight of moving clasts also generates pore water pressures, augmenting the buoyancy effects and lowering the shear strength of the debris. Reported values of water content fall between 10-30% by weight (Costa 1984).

- 5) internal stresses caused by particle collisions. The effect of a dispersive pressure causes upward lift of large clasts, which thereby become concentrated at the surface of the flow (Takahashi 1981).

The velocity of debris flows varies according to material properties (viscosity), track geometry, slope roughness, gradient and length. Reported values of velocity generally fall within the interval 0.5-20 m/s (Costa 1984, Table 1).

4.2.3 Initiating factors

The factors responsible for the release of debris flows are of two categories, external factors affecting a potential failure zone, and internal factors constituting the material properties and site characteristics of the failure zone. The complex interaction of factors involved in the initiation of debris flows is schematically illustrated in Figure 4.1, which is an attempt to account for important linkages, with no regard to relative weight. The initial failure is usually a slump, slide or topple. Flowage of the debris begins at varying distances of downslope movement of the material (Johnson & Rahn 1970, Costa 1984). The process of supply of water to a potential failure zone is an important factor in debris flow initiation. The external factor of overriding importance is heavy precipitation as rain (e.g.

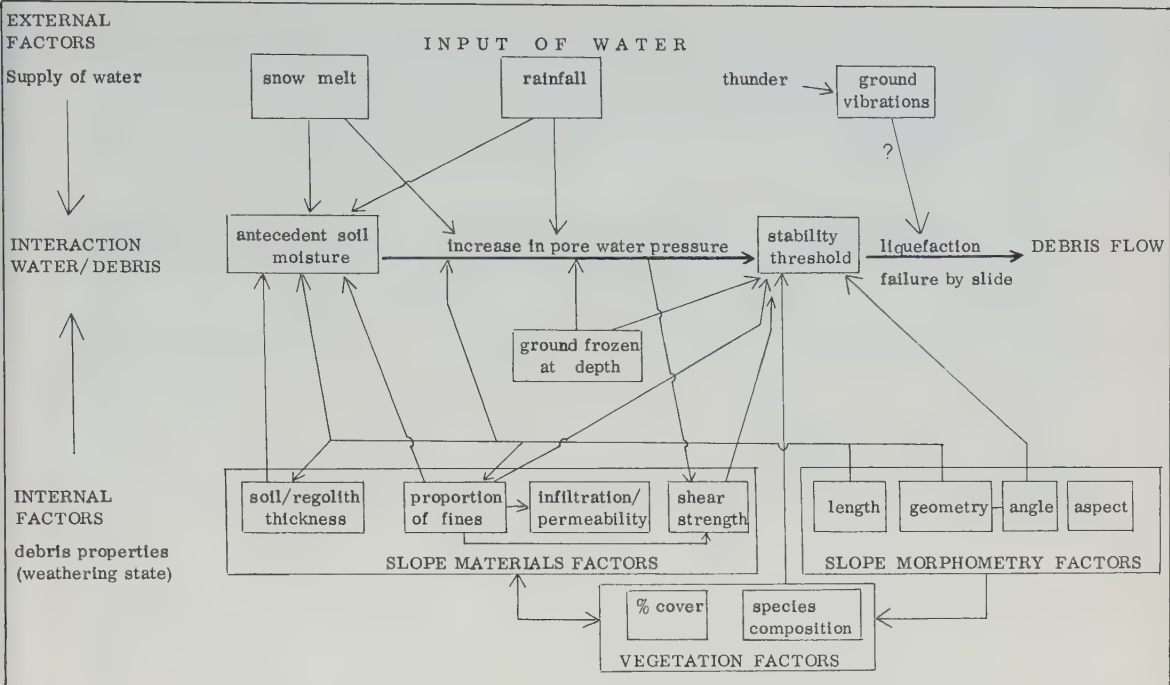


Fig. 4.1. General factor interaction in a slope system regarding debris flow initiation. The lower part of the diagram shows basic parameters of the system (material and site properties). The upper part displays imposed environmental stresses. In the middle section, the interaction of the two other levels is illustrated as changes in soil moisture and slope stability. See further text.

Strömquist 1976, Pomeroy 1980, Larsson 1982). Total amounts, e.g. in 24 hours, are not always possible to correlate to flow activity, although average values of rainfall leading to flow initiation may be estimated for different areas. Of more importance are short-term intensities, expressed as mm/hour. Prior et al. (1970) describe debris flow activity initiated at hourly intensities exceeding 6.3 mm in NE Ireland. Since rainfall intensities vary greatly locally, it is normally not possible to extrapolate meteorological records far from the recording station in order to estimate triggering amounts for observed failures (Op.cit.p.285). The relationship between rainfall intensity and duration leading to slope failures has been studied by Caine (1980). For short-lived rainfall of the convective type, lasting e.g. one hour, intensities of at least about 15 mm/hour are needed to release failures. Rapid snow melt may in less common instances supply the triggering amounts of water (Sharp & Nobles 1953, Owens 1974). A recent case where rapid snowmelt led to the release of a debris flow occurred on May 17, 1984 at Mt. Njulla near Abisko. A further external factor assumed to initiate debris flows are ground vibrations (Okuda et al.1980).

Internal factors involved in the initiation of flows are slope materials composition and weathering state, which may undergo a gradual change towards an instable situation. The triggering factor of most significance, strongly related to the external meteorological factors, is pore water pressure, illustrated on the intermediate level of Fig. 4.1. At high pressures, the shear

strength of the debris falls, increasing the risk of failure through lost cohesion between particles. Apart from the immediate input of large amounts of water as mentioned above, also moisture conditions during a preceding time period are important, when pore pressures may slowly build up to near critical levels for slope failure.

Flowage of debris may start by liquefaction, which can be described as 'a process that causes a saturated mass of soil suddenly to lose its shear strength and to behave like a fluid' (Brunsden 1979, p.139). The failure condition of a slope segment can be analysed using concepts from soil mechanics with the comparison of imposed stresses tending to initiate mass movement, and resisting strength properties of the slope material. These must be balanced if failure is to be prevented. This relationship is generally expressed by the safety factor, $F = \text{resisting forces} / \text{driving forces}$, where a value of less than 1 implies failure conditions. Pore water pressures can be seen to play an essential part in reducing shear strength of the debris if we look at the expression of strength derived by Coulomb:

$$S = C + (\sigma - u) \tan \phi$$

where S = shear strength of the debris, c = cohesion of the debris, σ = normal stress, ϕ = angle of internal friction of the debris, u = pore water pressure.

Normal stress subtracted by pore water pressure is usually referred to as effective stress, σ' .

4.2.4 Terrain localization

Debris flows may occur both along previous drainage lines and on slopes in-between, generally in steep terrain. A broad classification based on occurrence in the terrain is into hillslope flows and valley-confined flows, the latter following narrow valley bottoms (cf. Brunsden 1979 p.175). In some instances, this distinction may be difficult to retain since a hillslope flow can turn into a valley-confined flow in its lower course. A third category which is outside this general classification is the till flows which occur on glaciers (Lawson 1982). The starting zones for debris flows are usually located to spots in the terrain where water becomes concentrated, either by flow to concavities, or within the soil mantle by flowing along bedrock structures or on impermeable soil strata, e.g. a frozen layer (Strömquist 1976, Larsson 1982).

The depositional zones of debris flows are preferentially located either at the foot of steep slopes, where low gradient and lateral expansion of the flow with accompanying loss of water bring flowage to a stop (hillslope flows), or on alluvial or colluvial cones in front of valley or gully outlets (valley-confined flows). A description of typical terrain localizations of hillslope flows investigated in this study is given in section 4.3.

4.3 Debris flows in Nissunvagge. Distribution and geomorphological effects

4.3.1 The rainstorm-induced debris flows of 1979 southeast of Abisko. General description of the event.

A case of numerous debris flows released by a single intense rainstorm of unknown magnitude occurred on June 23, 1979 in the mountains south and east of Abisko. Several hundred mass movements were initiated on steep slopes, cf. Rapp & Nyberg (1981 a,b). Runoff following the heavy precipitation caused vigorous bank erosion and high sediment transport in streams draining the affected area, with flooding effects reaching more than 10 km from the centre of the area (Fig.4.2). The peak discharge of the Nissunjokk stream following the thunderstorm caused flooding of the fossil alluvial fan downstream of the canyon at the valley outlet (Fig.4.3). Further downstream, a large sediment plume extended into Lake Torneträsk from the delta of the stream Abiskojokk.

The large size of the flood in Nissunjokk was probably partly attributable to the direction of the storm track being downstream the Nissunvagge catchment, and the rounded shape of the basin, both factors giving simultaneous runoff entering the main channel from several tributaries (cf. Ward 1975, p.265-266).

The event in 1979 is by far the largest known case of rainfall-induced rapid mass movements in the Swedish mountains, comparable in size with for example the events at Johnstown, Pennsylvania in 1977 (Pomeroy 1980) and in Grand Teton, Wyoming in 1941 (Fryxell & Horberg 1943). It is one of several recently observed cases of extreme erosion caused by heavy rains in arctic areas, other events of simultaneous release of numerous debris flows being reported from Spitsbergen in 1972 (Larsson 1982) and 1981 (J. Åkerman, pers.comm.), and from Disko, west Greenland in 1984 (O. Humlum, pers.comm.).

The triggering meteorological conditions for the 1979 event are unfortunately only known in general outline, as the affected area lacks weather stations. Data from the lowland weather stations of Abisko, Torneträsk and Kiruna (cf. Rapp & Nyberg 1981b) are not directly comparable to expected precipitation amounts in the high mountain area, where the majority of mass movements occurred.

Meteorological situations with potential of causing rainfall-induced mass movements can empirically be divided into frontal passages with widespread rain (cyclonal rain) and local thunderstorms (convective rain). Around midsummer in 1979, northern Sweden experienced very warm temperatures, exceeding 20°C in the daytime (SMHI 1979), and connected with local thunderstorms, e.g. at Lainio associated with hail fall on the 23 June (total prec. 44.2 mm). The research station at Abisko recorded 13 mm rainfall in the afternoon the same day. The preceding part of the month had been dry with a total of 7.7 mm up till that day. The event of 1979 near Abisko was a violent thunderstorm during the late afternoon (15-16 h), with a probable precipitation of about 40-50 mm (Rapp & Nyberg 1981b). It seemed to move from southeast towards northwest and was probably most intense in the surroundings of the Nissunvagge valley (N-Å Andersson, Abisko, pers.comm.). As bedrock and soils in the mountains affected are fairly uniform (amphibolitic rocks, frost-riven debris mantles

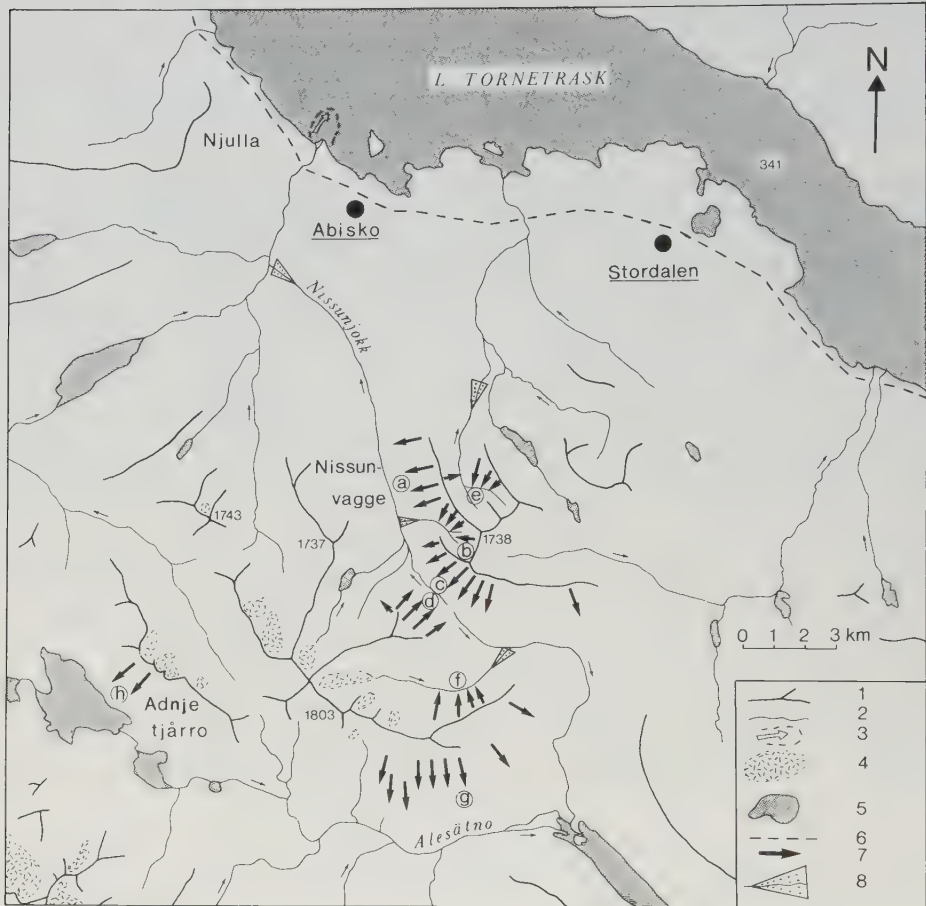


Fig. 4.2. Distribution of mass movements released by a rainstorm on June 23, 1979, in the mountains south of Abisko. Key: 1. mountain ridge, height in m a.s.l. 2. stream 3. sediment plume in Lake Torneträsk 4. glacier 5. lake 6. road and railway line 7. debris slides or flows (each arrow denote several cases) 8. coarse debris deposit formed by flood. Letters a - h refer to sections used in a survey of the mass movements (cf. Table 4.3).

and till of easterly origin), it seems likely that the distribution of mass movements triggered by the storm approximately outlines the zone of maximum precipitation (Fig. 4.2). Thus, it appears that an area of about 40 km² was strongly affected by high precipitation, with marginal effects outside this area. Isolated occurrences of mass movements most likely from June 23 are found at Mt. Rippatjåkka 2 km west of Torneträsk railway station, about 20 km from the Nissunvagge area eastwards, and on the southern slope of Mt. Adnjetjärro, some 10 km to the west.

4.3.2 Types of mass movement

Apart from the widespread effects by running water, the rainstorm of 1979 had a geomorphic impact on slopes in the Nissunvagge area by the rapid mass movements it released. These were of the following main types:



Fig. 4.3. Effects of the flood on June 23, 1979, on the fossil alluvial fan downstream the Nissunjäkk canyon (upper background). The existing main channel was considerably widened and several minor channels formed or reactivated during the event. Large amounts of material, ranging from silt to blocks in m-size, together with tree logs and other plant debris, were transported and deposited in the channel or on the banks. Air photo N.Å.Andersson, Abisko, 25.7.1979.

1. small debris slides, here termed short earth slides, released on till slopes of fairly low gradient and with a short transport range (cf. Fig. 4.4). The release scars were located to gelifluction lobes (turf-banked lobes).

2. debris slides released on steep grass-covered gelifluction slopes, where further transport of the debris was due to some debris flowage and very strong surface wash (cf. Fig. 4.5). The rounded slide scars expose bedrock. The scattered deposits of blocks, finer debris and slabs of the vegetation turf spread out on largely intact grass cover downslope indicate that the triggering cause of the slides may have been of the 'water blowout' type described by Temple & Rapp (1972, p.187). The range of these mass movements was about 200 m, although it is difficult to draw a distinction between the transport by mass movement and by running water.

3. debris flows, initiated as slides in sloping blockfields, rock wall ledges or chutes or on talus (Figs. 4.5, 4.6). These constitute the majority of failures during the event, transferring huge

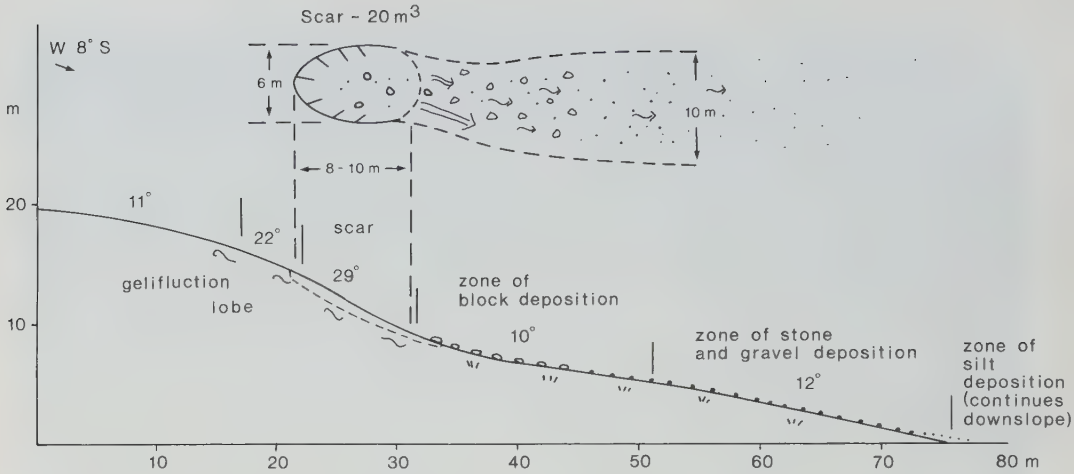


Fig. 4.4. Small debris slide ('earth slide') released at front of gelifluction lobe, section A, Nissunvagge. Note the sorting effect with respect to clast size occurring downslope of the scar, in response to fluvial transport of the slide material during and after the rainstorm. Short transport distance for coarse debris is typical of this type of mass movement, which is comparatively rare in Nissunvagge.

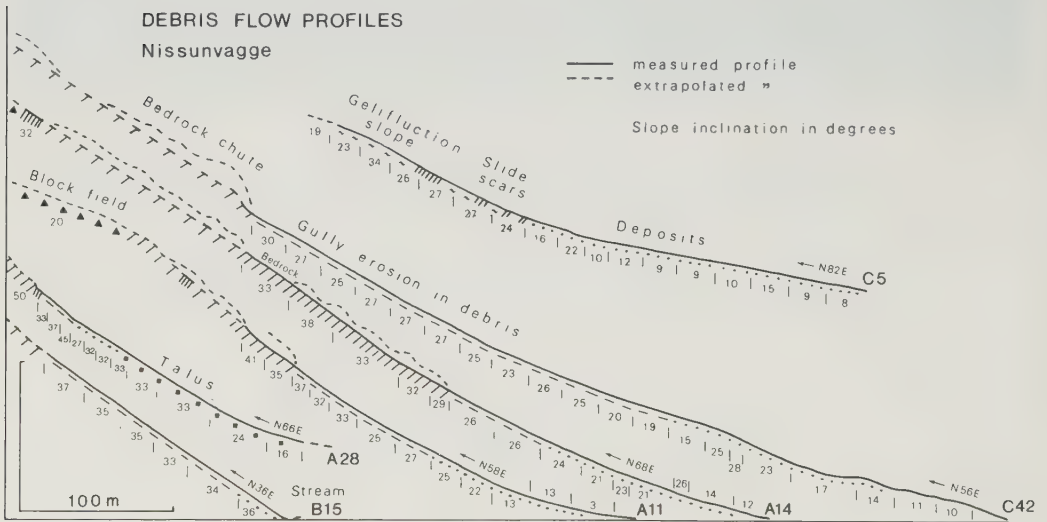


Fig. 4.5. Surveyed slope profiles along debris flows of 1979 in Nissunvagge. C5: debris slide with mixed transport mechanism on gelifluction slope (cf. text p. 45). A11, A14 and C42: debris flows released in rockwall chutes or blockfield above rockwall, and reaching foot of talus. A28: small debris flow released at junction of rockwall and talus, ending in upper part of talus. B15: debris flow released in rockwall chute, and reaching basal stream channel in V-shaped tributary valley.



Fig. 4.6. Debris flow no. C13, Nissunvagge, showing the characteristic features of most flows: a bedrock chute which served as channel for the flow in its upper parts, a gully with debris flow levees along the talus slope, and lobes of coarse, blocky material deposited at the foot of the talus, or on a grass-covered footslope. Approximate height of the talus 50 m. Photo 7.7.1980.

amounts of coarse debris downslope, but generally not reaching the local stream of the valley bottom. Their transport range was some hundred meters in general, with extremes of about 800 m. Triggering factor was presumably lowering of slope material strength due to increase in pore water pressures. The steeper slope gradients compared to type 2 facilitated the mobilization of larger masses of debris, which caused further erosion on their way downslope as debris flows.

4. combined debris flows/flash floods along small streams. These mass movements were released as slides in steep headwater areas, continued as debris flows along stream courses for much longer distances than the hillslope debris flows of type 3, or more than 1 km. They extended into the forest zone (Fig. 4.7). It seems likely that their anomalously long range can be attributed to particularly wet debris with high mobility caused by the torrential runoff of the streams. The erosion along the stream courses exposed the bedrock over long sections. Large amounts of material were deposited as outspread block lobes and silty sheets on slopes of only a few degrees inclination in birch forest.



Fig. 4.7. Very long debris flows reaching across gentle footslopes into forested areas near Alesätno river south of Nissunvagge. Aerial view towards west, southern slope of Mt. Pakkapahuktjåkka (cf. Fig. 4.9). Photo N.A.Andersson 31.7.1980.

A schematic illustration of the terrain localization of various geomorphic effects of the storm is given in Fig.4.8. Some further information on types and release conditions of the mass movements of 1979 in Nissunvagge was given by Rapp & Nyberg (1981 a,b).

4.3.3 Distribution of mass movements from 1979

A. Description

The general distributional pattern of the mass movements is shown in Fig. 4.2. In the survey of the event, the area was divided into several sections designated by the letters A-H. The detailed distribution of mass movements within sections A-D is included on Map I. Sections F-G are shown in Fig. 4.9. The remaining sections have not been mapped, only surveyed as to number of slides (Table 4.3). Section E is partly shown in the photograph of Fig. 4.12. The cartographic documentation regarding geometric form and dimension of the mass movements is not carried out in detail, as this would have demanded a more comprehensive field mapping project. However, inventory of all areas affected by mass movements during the rainstorm of 1979 was made to make possible future checks on new cases of slope failure and a follow-up of the long-term development of mass movements of known age.

The distribution of the debris slides and flows of 1979 shows a concentration within an area reaching from the southern slope of Mt. Pakkapahuktjåkka near Alesätno river, to the northern parts

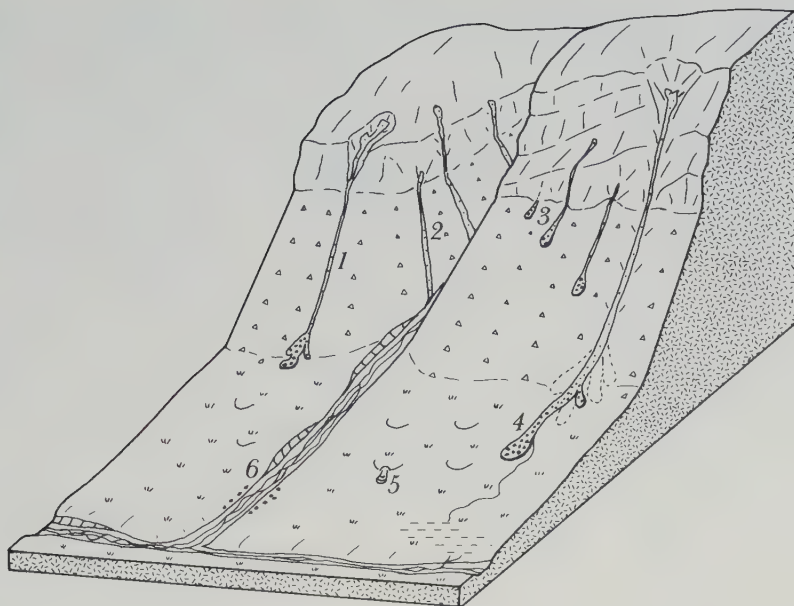


Fig. 4.8. Terrain localization of debris flows and other geomorphic effects of the rainstorm of 1979, Nissunvagge (schematic illustration).

1. debris flow released in depression or chute in mountain top blockfield or rockwall chute, with a range to the foot of the talus.
2. debris flow reaching local stream of v-shaped valley, where storm runoff removed the material further downstream.
3. small debris flows released in rockwall chutes or upper part of talus, accomplishing limited redistribution of material within the talus slope.
4. large debris flow released in chute, reaching out on gentle footslope beneath talus slope.
5. small 'earth slide' released at the front of a gelifluction lobe on a gentle footslope.
6. erosion in channel banks and overbank deposition caused by storm runoff.

of the Nissuntjärro massif. The localization of mass movements within this band in N-S direction, indicating the rainstorm track, is on slopes of between $31-40^\circ$ average inclination, especially from $31-35^\circ$ reflecting the angles of talus in the area.

The majority of slides started in steep rockwall sections and continued as flows across talus, often reaching the base of the scree slope and occasionally some distance across gentle footslopes of about 10° gradient. A large number of slides were released on steep sloping blockfields, which usually make up a convex slope unit above the rockwalls. In some instances, the new slide scars clearly contributed to a drainage network extension. Fewer slides had been initiated on gelifluction slopes of more gentle gradient. The relationships between number of mass movements, slope units and inclination for cases included on Map I are depicted in Fig. 4.10.

The largest number of mass movements had started at altitudes between 1100-1300 m. The lowermost deposits generally occur

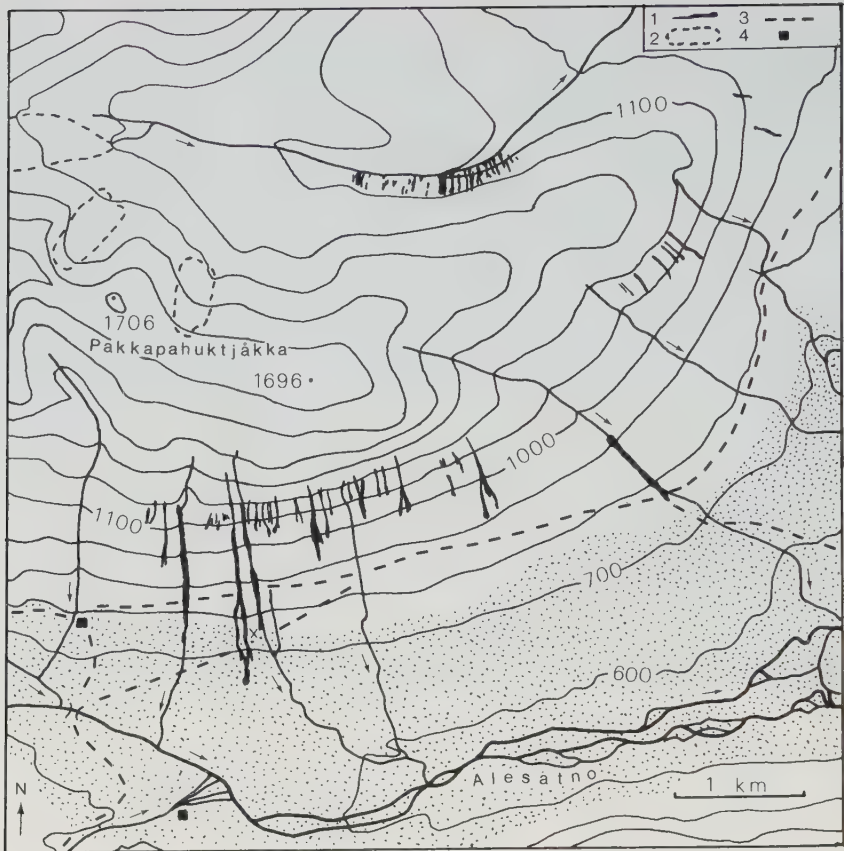


Fig. 4.9. Distribution of debris slides and flows from June 23, 1979, along the slopes of Mt. Pakkapahuktjåkka (section F and G, cf. Fig. 4.2.).

Key: 1. debris slide or flow 2. glacier 3. foot path 4. hut.

Area below tree line stippled. Heights in m. Position of old debris flow, dated by lichenometry, marked by x.

below 1100 m, down to about 900 m, but some small slides stopped at quite high altitudes. The variation in debris flow length has been estimated from Map I and field measurements and is shown, also related to slope units in Fig. 4.11. There is a positively skewed distribution towards small flows, although in comparison with e.g. cases reported by Owens (1972), Larsson (1982) or Innes (1985b) the sizes are large on the whole.

B. Evaluation of the distributional pattern

It is typical of convectional downpours to be highly variable in intensity, of short duration and seldom cover whole catchments (Starkel 1976, Caine 1976). Usually they affect areas of 20-50 km², lasting for about 0.5-1 hour (Barry & Chorley 1976). The spatial pattern of convectional storm rainfall is commonly elliptical, elongated in the storm track direction (e.g. Fogel & Duckstein 1969). The 1979 storm event appears to fit in well in these descriptions judging from its geomorphic impact.

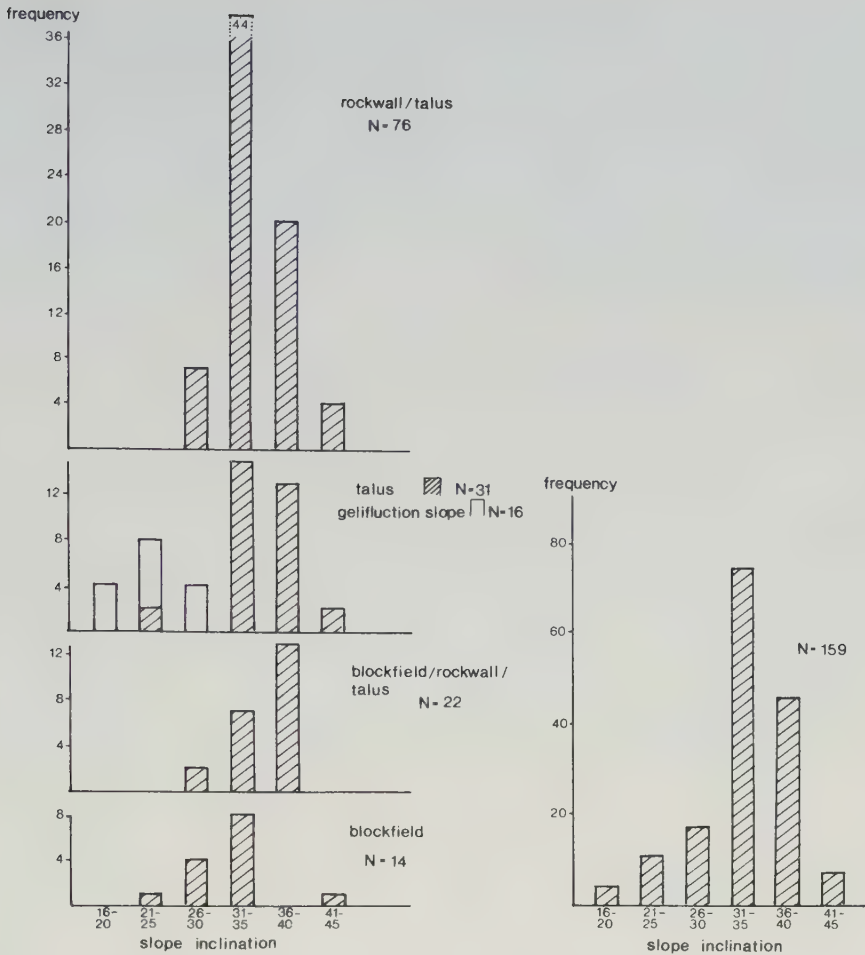


Fig. 4.10. Average slope inclination along debris flows in central Nissunvagge, for different affected slopes units (left) and for all slopes.

Slope aspect for the mass movements mainly mirrors the orientation of steep slopes in the area. Exposure towards the storm track direction for different rockwalls, a factor that might influence the intensity of local precipitation and thus be of importance with respect to mass movement release, has probably not been of notable significance, as slides were initiated also on the leeward sides with regard to the storm track in the southern part of Nissunvagge.

The lack of slides and flows on steep slopes on Mt. Pallentjåkka is most likely due to the local nature of the convective rain-storm. On the steep slopes of the E-exposed cirque of Mt. Nissuntjärro, the explanation for the lack of failures may also be that these slopes were still snow-covered by that time of the year, and the snow afforded a protection of the underlying slope by absorbing the rain and slowing down runoff and infiltration in the ground. Marked asymmetrical local distribution of debris flows occur in the V-shaped tributary valleys of sections B, E and F. This is treated in more detail in the following section. The debris flow occurrences at Mt. Rippatjåkka and Mt. Adnje-

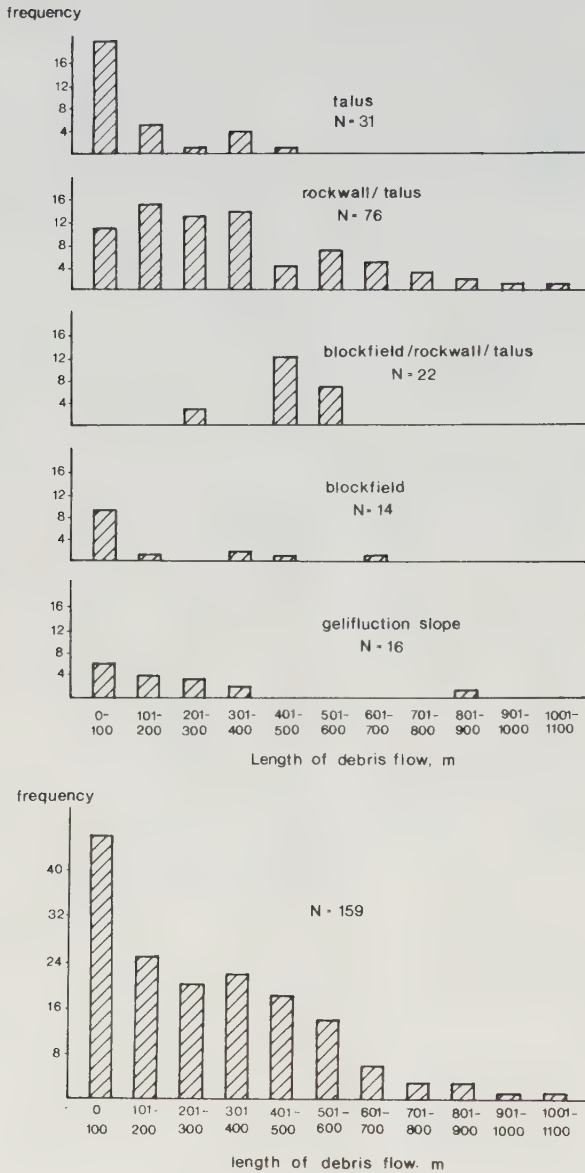


Fig. 4.11. Debris flow lengths in central Nissunvagge (approx. terrain lengths) for affected slope units and all slopes.

tjårro may be accounted for by assuming local thunderstorm cells of minor importance compared to the one over the Nissunvagge area. Isolated events at Torneträsk station in 1979 (Fig. 4.40) and later at Njulla in 1984 indicate that large flows may sometimes be initiated first, with smaller ones following only after continued rain as more stable terrain positions are mobilized. Triggering of small flows by the vibrations caused by an initial large flow is another possibility (Pomeroy 1980). However, other observations, such as in Nissunvagge 1982 (section 4.3.8), give evidence to the contrary, that is, initial release of small flows.

The erosive intensity of an event on the spatial scale may be expressed by the frequency of failures per unit distance or area. The average density of rapid mass movements from 1979 within the most heavily affected area is about $7/\text{km}^2$, although the concentration along steep slopes gives locally much higher densities, preferably expressed as no./km valley side. A comparison of spatial densities for this event with other cases of rainfall-triggered mass movements is given in Table 4.1. The densities in Nissunvagge greatly exceed the local densities of avalanche boulder tongues in some Lappland valleys ($2\text{--}5/\text{km}$), reported by Rapp (1960b, Table 19).

Table 4.1. Areal density and frequency of slope failures per km valley side in Nissunvagge and some other areas with rainfall-induced mass movement events

Locality and time of latest event	slide/flow no./ km^2	frequency no./km	reference
Nissunvagge 1979	7 (average)		
section A		8	this study
" B		26	
" C		20	
" D		14	
" F		25	
Longyeardalen, Spitsbergen 1972	18 (max)	12-15	Larsson 1982
Adventdalen area, Spitsbergen		about 3	Åkerman 1984
Kärkevagge 1959	1-2	3 (east side)	Rapp 1960b
Tarfala 1972	2	4	Rapp 1974
Coast Mts, SW Brit. Columbia, Canada	0.12	2-3	O'Loughlin 1972
Johnstown area, Pennsylvania 1977		25 (max)	Pomeroy 1980

4.3.4 Local asymmetric distributions conditioned by site factors

A very uneven distribution of debris flows from 1979 was observed in three small, partly V-shaped valleys, sections b, e and f (Fig. 4.2). The valley most strikingly affected by the rainstorm with respect to geomorphic impact is the one draining Mt. Nissuntjärro westwards (section B), cf. Map I. Possible factors influencing the observed asymmetrical occurrence of debris flows in this valley may be attributed to uneven rainfall distribution and runoff at the event, differences in relief and geomorphology, slope materials, vegetation and microclimate for the two valley sides. An attempt was made to evaluate the contribution of the various factors to the observed distribution of mass movements. Available time and equipment allowed only a general characterization of the conditions. A brief discussion of the possibility of uneven rainfall distribution between the valley sides is given in section 4.3.8.



Fig. 4.12. Asymmetric distribution of debris slides and flows in small, v-shaped tributary valleys of the Mt. Nissuntjärro massif. Upper photo shows section e, lower section b of Fig.4.2. In both cases, the mass movements are almost wholly restricted to the steeper valley side. Air photo Nils-Åke Andersson 31.7.1980.



Fig. 4.13. View of the V-shaped tributary valley (section B), looking downstream towards Nissunvagge. Note the marked asymmetry in cross profile, with the right (northern) side steeper than the opposite side. Photo 20.7.1982.

A. Relief and geomorphology

The valley is incised in the Nissuntjärro massif to a maximum of about 300 m in its central part. The cross profile is V-shaped with a marked asymmetry in steepness between the valley sides (Fig. 4.13). The valley head on the contrary is rounded, cirque-like and probably fashioned by glacial erosion (cf. Maps I and II).

The partly southfacing valley side is the steepest downstream of the valley head and characterized by a rockwall dissected by chutes in the upper part, and talus at the base, reaching the main stream channel. Some rock outcrops occur adjacent to the stream. Above the rockwall commences an extensive blockfield.

On the opposite side of the valley, the slope consists mainly of till with gelifluction and patterned ground features (stone stripes). The profile is slightly convex and the steeper, lower-

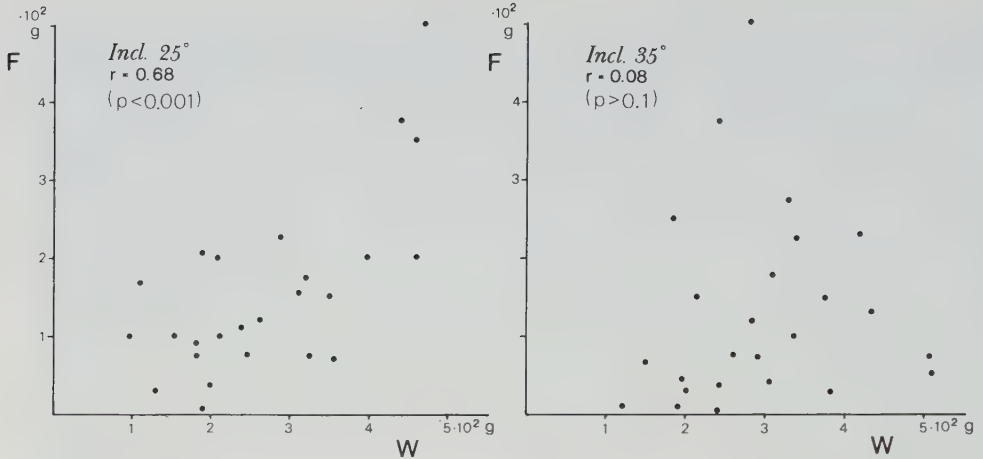


Fig. 4.14. Relationship between shear force (F), needed to dislodge particles downslope, and weight of particles (W) on the opposite slopes of asymmetric tributary valley, Nissunvagge. On the steeper slope, no positive relationship between force and weight was found, indicating a generally high instability even for large particles. Measurements were made using a spring balance. Both F and W are expressed in grams. p =prob.value.

most part is locally affected by slides or short debris flows. As no equipment for geotechnical stability evaluation of the slopes was available, a simple method was used to compare the conditions on the two sides. To quantitatively evaluate the difference in slope angle with respect to surficial material stability and transfer, measurements of shear forces needed to dislodge particles in a downslope direction were undertaken, using a spring balance. The results showed a positive correlation between particle weight and applied force for the northfacing side (25° inclination), whereas on the opposite, steeper side (35°) no relationship was evident (Fig. 4.14). This implies that even large particles were in unstable positions near their angle of repose on the steep slope, and therefore easy to move also with low applied force. On the slope of 25° gradient, the clasts were generally in more stable positions.

Relative height and size of contributing areas to runoff are larger for the southfacing valley side, factors that apart from the steeper slopes, might be evoked to explain the prevalence of mass movements on this side. A morphological factor of possible importance in this context is the chute sculpture, giving a higher potential for runoff concentration compared with the opposite slope.

B. Slope materials

On the steep south-facing side with rockwalls, the till is hidden beneath a thin blanket of talus derived from the local bedrock (amphibolite). The till cover is between 0.5-3 m thick with

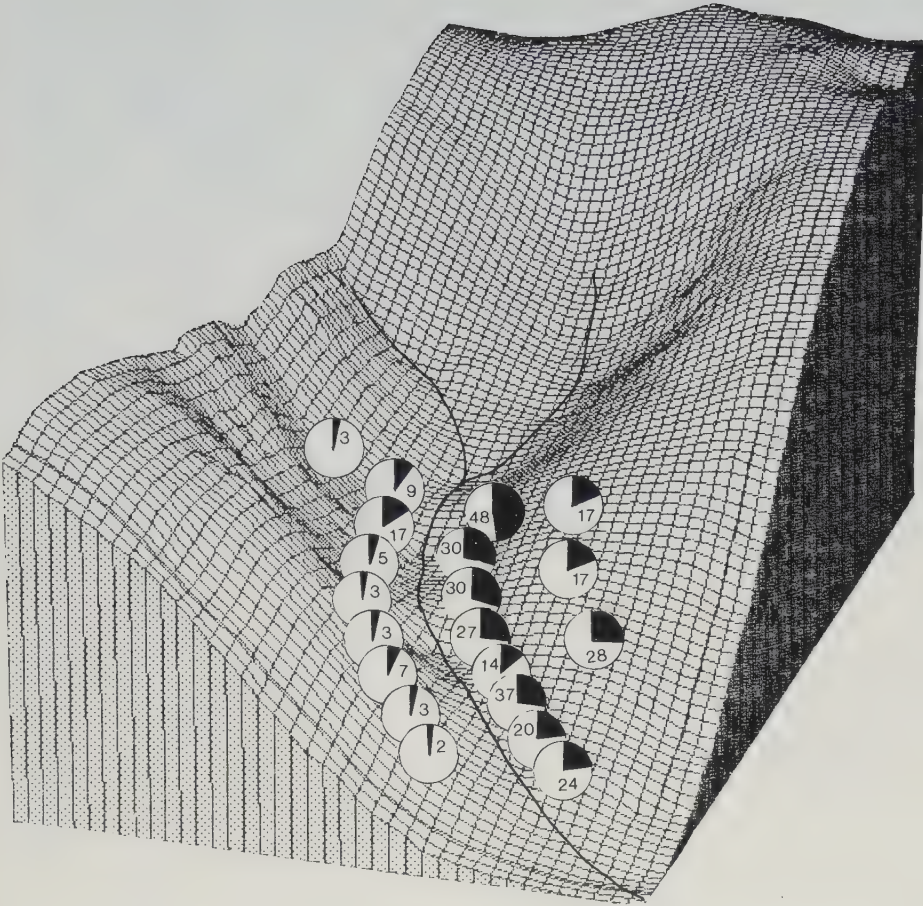


Fig. 4.15. Percentage of erratics on the opposite slopes of section B, Nissunvage. The left-hand slope (southfacing) consists mainly of local talus material (amphibolite), whereas the opposite slope is covered by till with many erratics of easterly origin (granites, gneisses a.o.). Length of valley about 2 km. Total number of erratics found on southfacing side 55 ($n=992$) or 5.5 %, on northfacing side 312 ($n=1181$) or 26.4 %. All differences between pairwise measurements statistically significant at the 5% level or less (chi-square test).

increasing depth downslope, judging from sections created by the many debris flows in 1979. On the north-facing slope, till dominates, with unknown thickness. Material contrasts for the two slopes are indicated by a sampling of percentage of erratics (Fig. 4.15) and grain size (Fig. 4.16), showing a slightly finer and more exotic composition of the material on the northfacing slope. Soil moisture and type of vegetation therefore differs between the valley sides, the southfacing side generally being the driest. Water content varied between 5-17% of dry weight according to samples taken during the study period. An attempt to

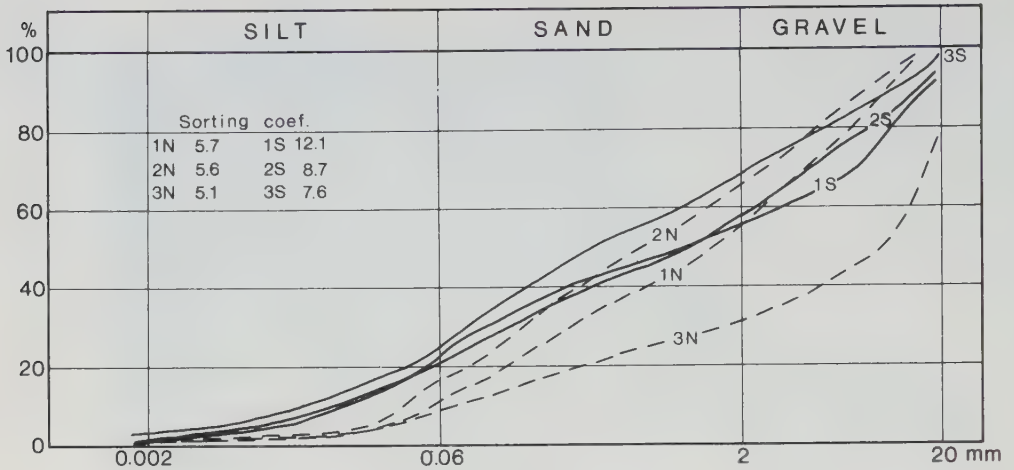


Fig. 4.16. Grain size distribution of soil samples from the opposite slopes of the tributary valley, section B, Nissunvagge. N= northern side (southfacing), S= southern side (northfacing).

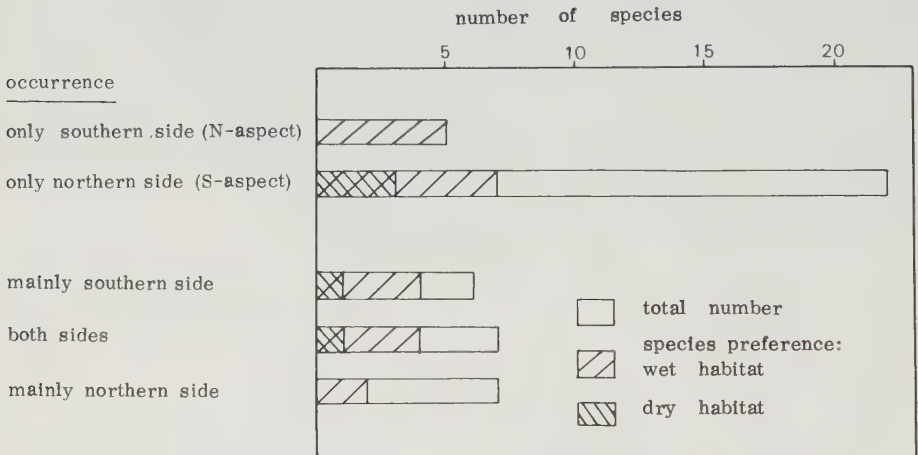


Fig. 4.17. Number of plant species (excl. lichens and mosses) found on the valley sides of section B, Nissunvagge. The southfacing side has a higher total number of species than the northfacing, which is dominated by a few moisture-adapted species. Cf. appendix 1.

monitor soil moisture periodically with the use of buried gypsum blocks, which can be read with resistance meters, failed, perhaps because the blocks used were not suited to the wet conditions obtaining on the studied slopes. Weather conditions also prevented water infiltration measurements from being made.

C. Vegetation

Differences in insolation, soil moisture and persistence of snow cover between slopes of north and south aspects are factors that

significantly influence the vegetation, both regarding species composition and percentage ground cover. From the viewpoint of slope stability, the most important factors related to the vegetation are probably the cohesive strength in the upper soil layers afforded by root penetration and the protection of the soil surface against direct raindrop impact and surface wash erosion. These and other factors have been thoroughly studied in investigations on soil erosion, though not very often in arctic/alpine areas.

In general, the vegetation of the small stream catchment B is discontinuous and dominated by grasses or meadow-type species on the south-exposed slope. The lower parts of the slope have locally dense field cover, in spite of very steep gradients. The north-exposed slope is characterized more by moisture-adapted herbs and a low percentage cover, lacking a continuous grass turf.

A comparison of species diversity with respect to preferred habitats (dry vs. humid) for the two slopes is shown in Fig. 4.17. A listing of the species found appears in Appendix 1. The qualitative observations made provide some measure of the contrasts in vegetation between the valley sides.

D. Microclimate

Contrasts in microclimate between opposite valley sides may be pronounced and both directly or indirectly, through vegetation or soil moisture, influence slope processes (e.g. Geiger 1965, Churchill 1982). Some limited measurements of ground and air temperatures near ground level were carried out for the two valley sides. Measurement points were chosen along three cross-profiles of the valley. Temperatures were measured at 0.5 and 1 m depth using thermistors encapsulated in plastic tubes, and at depths of 20, 10, and 5 cm and at the surface using ground thermometers. Furthermore, max-min-thermometers were placed on the ground close to the other thermometers. Readings were made daily between 10-11 h during the periods 14-18 July and 21-23 July. These data may serve to give an indication of the magnitude of differences on a short-term basis between the two valley sides.

Ground temperatures.

The measurements showed that no frozen ground remained even at 1 m depth. The south-exposed side was several degrees warmer than the shady side at this depth. Also at 0.5 m depth, a difference of more than 2 degrees usually prevailed throughout the readings. Vertical temperature profiles in Fig. 4.18 show this difference during various days. The progressive warming up of lower levels with time can be seen, as also the quicker response to prevailing weather conditions in the upper layers. Diurnal temperature profiles based on four measurements during a day are compared for the opposite sides in Fig. 4.18. The warming up of surficial layers is marked at 1400 hours and reaches about 5 cm depth below which temperature changes more slowly. Maximum temperature contrast at the ground surface between the valley sides occurred at this time and exceeded 6 degrees. Contrasts were smaller below the surface, or about 0.5-2.5°C.

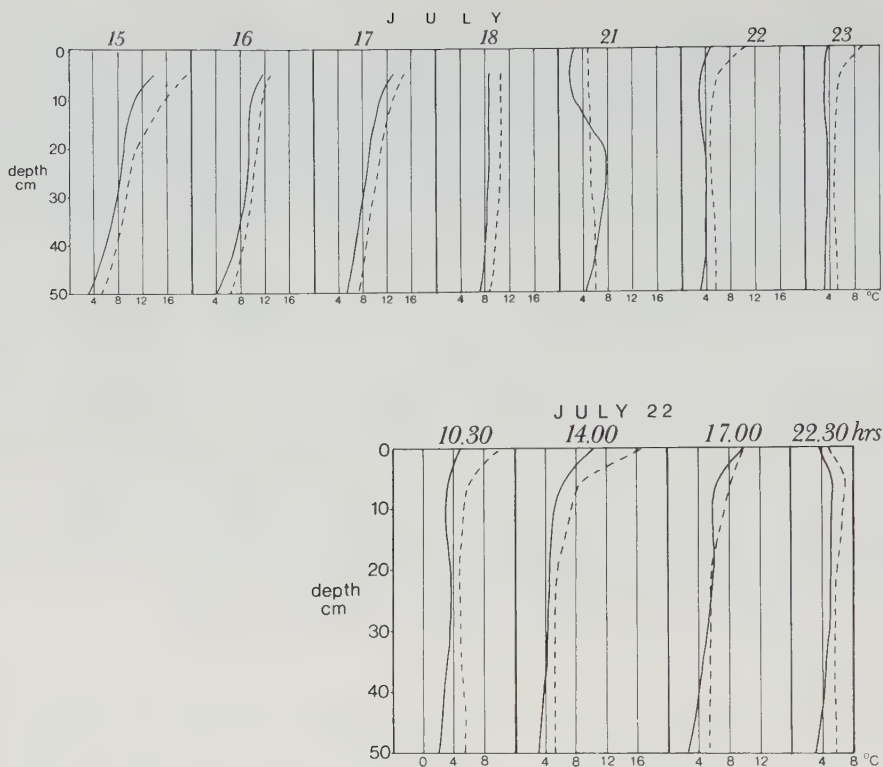


Fig. 4.18. Ground temperatures during the period July 15-23, 1982 (above) and diurnal variation in ground temperature on July 22, north- and southfacing slopes of section B, Nissunvage. Dashed line= southfacing slope. Cf. text.

Air temperatures.

Air temperatures on the slopes near surface level were compared with temperature records from a standard weather screen located near the outlet of the valley. Measurements were initially made with shielded thermometers (14-18 July), subsequently without shielding (21-23 July, 19, 21 August). The screen also had a THG running during most of the summer, with monthly restarts (Fig. 2.8). As expected, contrasts between the valley sides were even more pronounced than for the ground temperatures. Maximum temperatures on the southfacing slope frequently exceeded screen maximum temperature by more than 15° , and northfacing slope temperature by more than $9-10^{\circ}$. Minimum temperatures were slightly lower on the southfacing slope than screen minima, and generally slightly higher than minima of the opposite slope. The data bring out the more extreme range in temperature conditions for a south-exposed slope compared with a shady north-exposed slope, as well as for sites near ground level compared with a standard screen at about 1.7 m (Table 4.2).

E. Concluding remarks

There are obvious differences between the two valley sides concerning steepness, morphology, materials, vegetation and

Table 4. 2.

Daily extremes in air temperature on north- and south-facing slopes, section B.

day	maximum temperature									
	site:	north-facing		south-facing		north-facing		south-facing		
		A2	A3	A5	A6	C12	C13	C15	C16	screen
July										
14		16.0	15.5	31.9	28.0	-	-	-	-	10.7
15		19.5	16.5	38.0	34.7	-	-	-	-	18.7
16		24.0	17.3	27.5	34.6	20.2	18.0	24.3	28.9	18.5
17		19.1	19.1	29.0	31.3	20.2	18.3	31.3	25.5	19.3
18		21.7	19.0	28.9	27.1	19.7	12.9	25.0	24.0	14.5
21		10.6	8.7	3.5	2.7	8.0	6.0	8.6	10.5	3.5
22		13.0	10.4	18.8	-	8.9	6.2	14.4	10.4	3.0
23		14.9	12.1	18.0	-	10.3	6.7	15.0	13.1	5.0
Aug										
19-21		11.7	11.2	18.6	21.0	10.0	-	17.0	15.9	9.0
		minimum temperature								
July										
14		0.0	-0.8	-1.2	-1.1	-	-	-	-	0.3
15		0.6	-0.4	3.0	4.3	-	-	-	-	6.3
16		12.0	-	12.2	(4.3)	10.5	9.9	12.0	(0.9)	12.6
17		11.9	11.7	11.8	11.8	10.3	10.4	11.2	11.2	11.9
18		7.0	6.8	7.7	7.0	6.1	6.5	5.7	7.4	6.2
21		0.4	0.0	2.7	0.2	-0.6	-0.4	-2.5	-0.2	-1.0
22		0.4	0.3	-0.2	-	-0.9	-0.7	-2.8	-0.8	-0.6
23		0.2	-0.3	0.8	-	0.3	0.3	-1.3	-0.4	-0.5
Aug										
19-21		5.7	5.0	5.5	5.2	5.0	5.0	4.0	5.0	4.5

microclimate.

Without further investigations of a geotechnical nature it is not possible to draw more than tentative conclusions on the role of the various factors in the localization of the debris slides and flows of 1979.

1) The aspect factor may be regarded as less important, taking also the other two valleys with asymmetric debris flow occurrence in consideration. The flows were located to a north-exposed slope in one case, contrary to the other two.

2) Differences in material and vegetation are difficult to evaluate. Possibly, the finer material, more easily mobilized by high pore water pressures, and low vegetation cover on the northfacing slope may speak in favour of this side as the most erodible. Till slopes were affected by slides elsewhere in Nissunvagge during the storm (cf. Maps I and II), indicating that material factors are not primarily responsible for the asymmetry in slide occurrence.

3) Valley side contrasts in gradient and surface configuration (morphology and extent of upper slope segments, favourable to concentration of runoff) are factors that all three valleys have in common. It may be concluded that these are the most probable factors responsible for the asymmetric distribution of mass movements in the valleys.

4.3.5 Erosion and transfer of debris by the 1979 event

The geomorphological significance of an erosional event can be stated either in terms of landform changes (relief-forming effects) or as accomplished denudation (slope lowering) (Starkel 1976).

In the event southeast of Abisko, several drainage basins of different orders were affected. As no measurements of discharge or sediment yield are made for the streams, the amount of sediment output from the area struck by the rainstorm cannot be estimated. However, air photos and field observations show that much of the eroded material did not leave the basins but was merely transferred to the base of steep slopes and stored as new landforms (isolated debris flows and compound debris flow cones). Thus seen on the basin level relief-forming effects were more important than denudational.

The air photo survey made of the event and field measurements of the size of flows of various scales allows us to make a rough estimate of the masses involved in internal transfers within the basins. To estimate the total amount of mass, field measurements were made of ten debris flows, representing different size classes, and the volumes of the deposits plotted against horizontal lengths of the flows (Fig. 4.19). A rank correlation revealed a significant positive relationship between volume and length. A curve fitted by eye was used to graphically find the volumes of other debris flows, the lengths of which were measured on Map I. The total mass of debris deposited in debris flows of 1979 was calculated roughly to 134 000 m³ (Table 4.3). Allowing for a pore volume of the debris of 30%, a value of 97 000 m³ of eroded rock material is obtained. Calculated for the drainage basin of Nissunjokk, a momentary erosion of the order of magnitude of 72 000 m³ or 1.9 mm is estimated to have occurred due to the rainstorm. The actual value is probably higher, as the amount of material exported from the basin by the stream is not included.

The western tributary valley of Mt. Nissuntjärro makes up a subcatchment (2km²) where a somewhat more reliable value of local denudation can be obtained. Almost all of the debris flows of this valley reached the main stream channel and the material was largely exported from the catchment by the flood waters following the storm. Heavy fluvial reworking occurred of the channel downstream of the valley outlet, with formation of large boulder levees 1.5 m high. Only marginal widening of the channel took place, indicating that its size was already adjusted to large storm events, periodically also influenced by slush avalanches (cf. Fig. 6.2).

An estimate of the peak discharge was made from a surveyed cross profile between high water marks visible on the channel margins near the outlet of the valley. The channel gradient is about 8°, width 20 m and maximum water depth during the flood about 2 m. Using the Manning equation (cf. Embleton et al. 1979) velocity of the flow is

$$V = \frac{1}{n} R^{2/3} S^{1/2} \quad \text{or about } 7.5 \text{ m/s,}$$

where V =mean channel velocity, R =hydraulic radius (channel cross-section area/wetted perimeter), S =water surface gradient, n =resistance coefficient (about 0.05 for this type of channel roughness, cf. Simons 1969, Table 7.1.2).

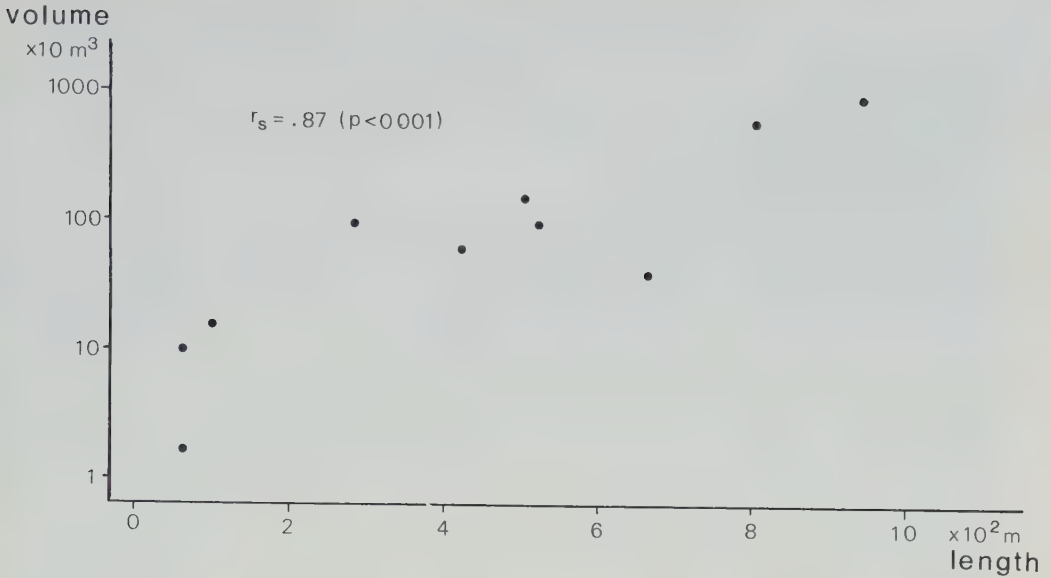


Fig. 4.19. Relationship between volume and length of debris flows in the Nissunvage area. Based on field measurements.

Table 4.3. Mass of displaced slope debris by rapid mass movements during the rainstorm event on June 23, 1979. Estimate based on measurements of deposits in the field and on air photos.

field section	debris volume, m ³	no. of d.flows	average volume, m ³
A	7000	31	225
B	17000	41	415
C	67000	97	690
D	12000	34	350
E	1200	20	60
F	7300	32	230
G	20000	42	490
H	2500	5	500
total	134000	302	440

This velocity is probably well above what is needed to cause the observed channel modification downstream the valley outlet (cf. Bradley & Mears 1980). This gives a value of the order of magnitude of 150 m³/s for the peak discharge, which is probably an overestimation. As the period preceding the storm event had been comparatively dry, it is probable that the slopes possessed a considerable storage capacity for rainwater, a fact which further supports the likelihood of an extremely high intensity of the rainstorm, causing the indicated runoff and observed erosional effects. The large magnitude of the runoff from the valley is evident when comparing with catchments of similar size affected by floods after heavy precipitation, e.g. those described by Russell (1972) from British Columbia, where peak discharges of 100-115 m³/s were estimated. However, the high water marks in

Nissunvagge may have been formed during a temporary surge. Irregularities in discharge may occur if a channel is blocked by snowbanks or by debris flow material.

Total amount of transported debris is estimated at about 17000 m³ corresponding to a momentary rock denudation of 5 mm in the subcatchment.

4.3.6 Landforms created by the 1979 event

Rapid mass movements such as debris flows cause a local change in slope morphology. The resulting landforms can be separated into erosional and depositional.

A. Erosional forms.

Debris flows are associated with two erosional forms, rounded or elongated release scars, and linear channels or furrows formed by the flow.

Release scars

As stated previously, debris flows often start as slides. Whole sheets of coherent material start to glide downslope over a flat or gently curved slip surface, leaving behind a generally rounded scarp and a depression downslope, often with exposed bedrock (cf. Fig. 4.20). In Nissunvagge, the slide scars range in size from a few meters to about 20 m in width and 0.5-2 m in depth. The inaccessibility of most scars on steep rockwall ledges or in chute sidewalls have precluded closer inspection.

Debris flow channels

Downslope of the scar, movement changed into flow through the loss of cohesion of the material. Debris flowage often erodes a several hundred meters long channel which tends to be U-shaped (cf. Johnson 1970) and flanked by debris levees. However, some flows accomplish very little erosion below the release scar, lacking well defined channel and actually flowing on the substrate with little damaging effect. The debris flow channel often follows old drainage lines or otherwise adjusts to the local topography of the slope. Frequently the channel is split into two arms in the lower part of the flow. A large percentage of the longer flows are of this 'forked' type (Figs. 2.5, 4.24).

In Nissunvagge, debris flow channels are generally incised into till or talus material to depths of 0.5-5 m and widths of 2-10 m. Frequently bedrock is exposed in the channel bottom. The general association of debris flow channels with levees (see below) sets them apart from stream gullies or other linear depressions trending downslope.

Erosional forms are replaced by depositional when the roughness of the slope increases, e.g. due to the presence of large boulders, or with decreasing gradient towards the slope foot.

B. Depositional forms

Two basic forms are distinguished, levees and lobes. These may build larger, cone-shaped features, resembling alluvial cones but of much coarser material and lacking stream channels. These large depositional forms are referred to as debris flow cones (cf. Rapp & Nyberg 1981b).



Fig. 4.20. Release scar of debris flow A11 Nissunvagge. The flow started in a shallow depression of the sloping blockfield of Mt. Nissuntjärro, leaving a one m deep and about 10 m wide scar. It exemplifies the common association between slope concavities collecting runoff, and slide initiation. The steeper convex slope in the background remained stable during the rainstorm. Slope gradient at the scar 32° . Photo 7.7.1980.

Levees

Debris flow levees may occur both along the erosional part of a channel, and after erosion has ceased, with a preserved ground surface in between. They are thus formed as immobilized parts of the flowing mass, and not as material that has been plowed sideways on location by the passing flow. Their size in Nissunvagge varies from gravelly ridges of a few dm, to blocky ridges 2 m high and about 5-6 m wide. Often levee formation is discontinuous along a channel, and at bends the outer levee can be considerably larger than the inner. The characteristic cross profile of levees is asymmetric, usually with the steepest side facing outwards from the channel. Unless partly erased by avalanche erosion, debris flow levees are very distinctive forms and the best diagnostic of debris flow activity on talus slopes. On slopes with features like block streams or large sorted stripes, levees may be less easy to distinguish, especially if no channel is developed.

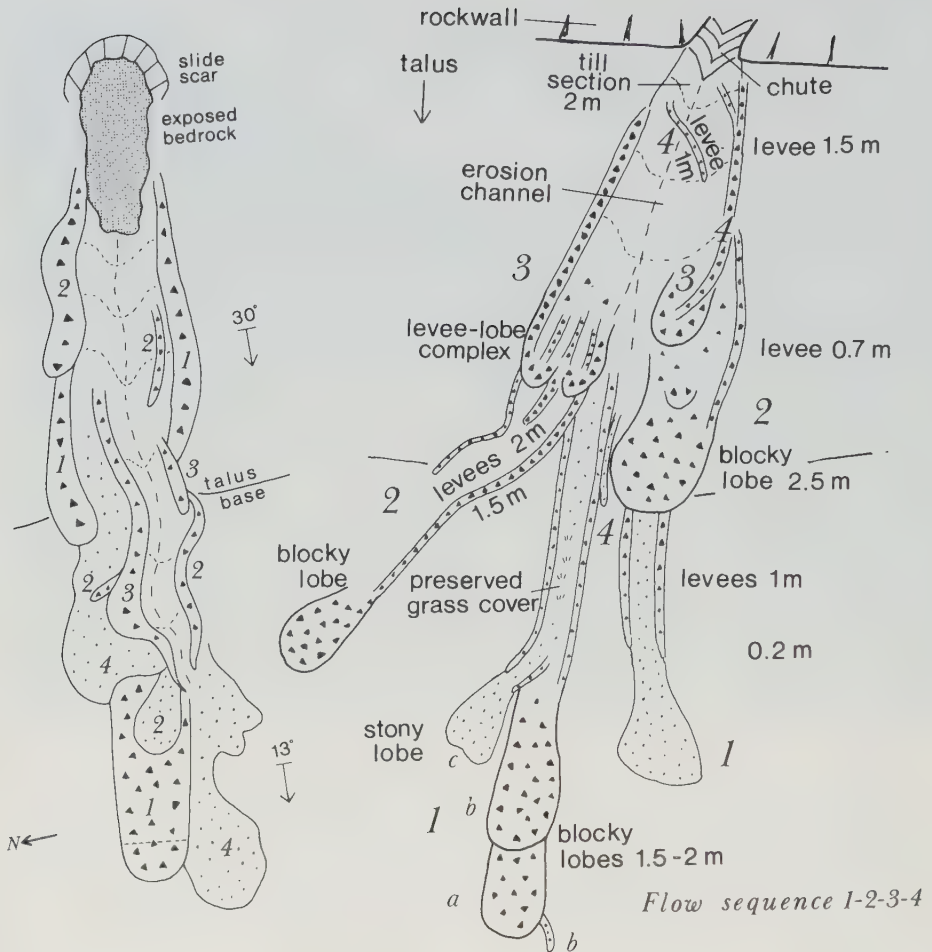


Fig. 4.21. Schematic maps of debris flows A4 and A12, Nissunvagge, showing configuration and flow sequence of levees and lobes. Not to scale. Approximate length of A4 250 m, A12 350 m. Values in m denote height or depth of features in relation to surrounding ground surface.

Lobes

Debris flow lobes generally occur at the lower end of a channel, where motion has ceased due to decreasing gradient. Local obstructions cause lobe formation also on steep slopes. The size of lobes in Nissunvagge ranges from miniature features of sand and gravel in dm-size, to blocky 10-15 m wide and more than 50 m long accumulations.

The occurrence of successive pulses can often be induced from cross-cutting relationships between levees and lobes. This is exemplified by the two cases shown in Fig. 4.21. Reconstruction of the temporal sequence of up to four different surges seems possible from the morphology of overlapping levees and lobes. The observations indicate an initial flow surge of coarse debris, which is the source of the largest and lowermost situated lobe,



Fig. 4.22. Convex front of debris flow lobe A4, Nissunvagge. View towards the south. The white line across the lobe was painted in 1980, as a field experiment to monitor subsequent movement of the material. Almost no movement occurred within the period 1980-1984, indicating that the debris flow accumulation has retained its core of fine material, and has not caused any marked depression of the underlying sediments following initial deposition. Slope gradient 13° . Photo 17.7.1980.

followed by one or several smaller flow surges of less coarse debris, which is deposited laterally to or partly overlapping the first coarse lobe. Some of these lobate deposits are well sorted and may be water-laid, of the 'sieve-deposit' type (Hooke 1967). Finally, fine material is washed out of the channel by a more fluid phase to lower elevations in front of the debris flow lobes, forming fine sheets of sorted sand and silt.

Debris flow lobes in Nissunvagge either form single terminal accumulations or several deposits adjacent to each other below the channels. The morphology of the lobes seems to depend on material properties, notably clast size distribution of the flows, and on the topography and gradient of the depositional zone. Coarse, blocky flows, especially when deposited on gentle footslopes, form rounded convex lobes (Fig. 4.22). The front curvature of such lobes has been shown to depend on the strength properties (viscosity) of the debris (Johnson 1970). Finer, stony or pebbly flows, especially when stopped on steep slopes, form flat, tapering lobes (Fig. 4.23). The absence of rounded convex front may be related to lower strength (viscosity) or to secondary collapse of the flowlobe as an adjustment to a stable position on the steep slope (cf. Takahashi 1981). The great kinetic energy of large flows is one factor contributing to their long range. Another factor may be that they often follow pre-existing gullies. Small debris flows may come to rest on steeper slopes as the mentioned factors are less important influences on their motion. Quantitative data for a number of flows regarding the degree of branching and channelling as well as the inclination of terminal lobes illustrate the relationships (Fig. 4.24).



Fig. 4.23. 'Tapering' debris flow lobes in section F, S. Nissunvagge. Notice the clear contrast between the old, dark, lichen-covered block slope and the fresh, light-coloured material of the flows. Photo 16.7.1980.

Old debris flow lobes may be difficult to distinguish from block streams or stone-banked gelifluction lobes, if no clear connection to levees is evident.

Cones

There are two basic types of cone- or fan-shaped depositional landforms in Arctic mountains, those due to fluvial deposition (alluvial cones) and those due to various mass movement processes. The widespread importance of mass movements other than rockfall in shaping the latter forms is nowadays recognized,

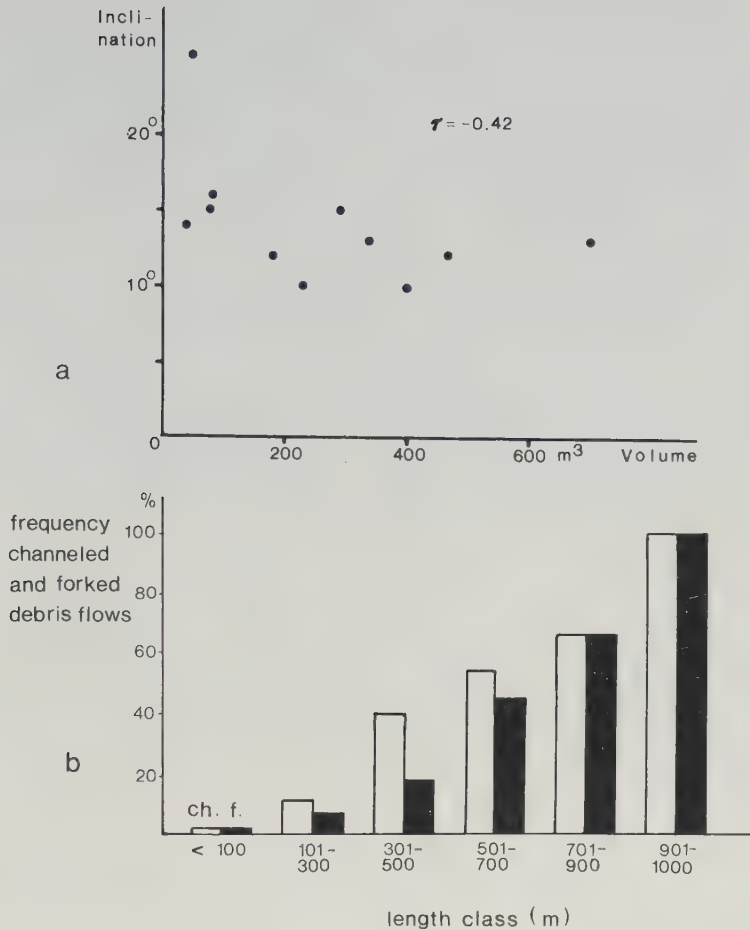


Fig. 4.24. a. Relationship between slope inclination and size of terminal lobes of debris flows, section C, Nissunvagge. Weak negative correlation shown by Kendall's tau ($p=0.036$). b. Increasing tendency of larger debris flows to be channelled and develop branches in their lower reach.

and the terminology has been extended to account for the various process influences, although different authors use different terms (e.g. Rapp 1959, Schweizer 1968, Luckman 1977, Church et al. 1979, White 1980). As we often find, nature resists our attempts at classification by allowing intermediate types to exist, with gradational differences due to relative process activity.

The various types of accumulations occurring in the Nissunvagge area are depicted on Map II, and four cases compared with regard to slope profile in Fig. 4.25. The distinguishing features of debris flow cones are gradients intermediate between rockfall and avalanche talus, and superimposed generations of levees and lobes of different age, or representing different surges of a single flow. Apart from a channel near the apex, scoured by repeated flows, drainage is below the surface of coarse blocks. An example of a large debris flow cone is no. C42, central Nissunvagge (cf.

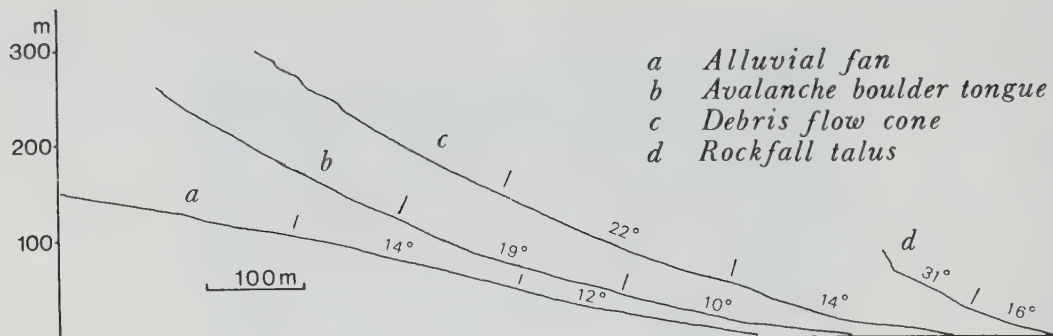


Fig. 4.25. Four types of debris slope accumulations in Nissunvagge. Average slope gradients given for upper and lower parts of the accumulations. The debris flow cone shown is intermediate in gradient between the rockfall talus and the avalanche boulder tongue. Fronts of individual flows (lobes) give locally steeper gradients on the cone, not shown in the profile.

Fig. 4.39 and Rapp & Nyberg 1981b). This cone has several generations of older levees and lobes with clearly different lichen covers (cf. section 4.4). Rockfalls and also snow avalanches have a certain impact on the cone, but debris flows are responsible for the morphology at large.

C. Influence on slope morphology

Avalanches and debris flows are important processes causing a modification of rockfall-generated talus in periglacial areas (e.g. Rapp 1959, Bones 1973, Luckman 1977, White 1980). Their relative significance in any particular area can often be assessed from an analysis of the prevailing morphology or material sorting of debris slopes (Kotarba & Strömquist 1984).

Morphometric relationships between debris slopes and other slope units in Nissunvagge were investigated by measurements on the contour map I. Bearing in mind the limitations of scale (1:10 000) and absolute accuracy of the map (ch. 3), only rank correlation was used when studying the association of different variables (Fig. 4.26). The data indicate significant correlations especially between the scale of debris slopes, rockwall height and occurrence of chutes. The longest debris slopes tend to be less steep and with a more marked basal concavity than shorter ones. Basal concavity and plan convexity are associated with the occurrence of large chutes and debris flow cones. As debris flows and avalanches have longer reach than rockfalls, they tend to extend the talus into lower footslopes. Simultaneously, especially snow avalanches may suppress the upper limit of talus on rockwalls, as noted by Rapp (1960b). This effect is noticeable on the rockwall of section D (cf. Map II). It should be pointed out that the positive correlation between slope length/height and rockwall height expresses the static situation presently, not the dynamic relationship between the variables. Upwards growing debris slopes should theoretically lead to a decrease in rockwall height with time by slope replacement.

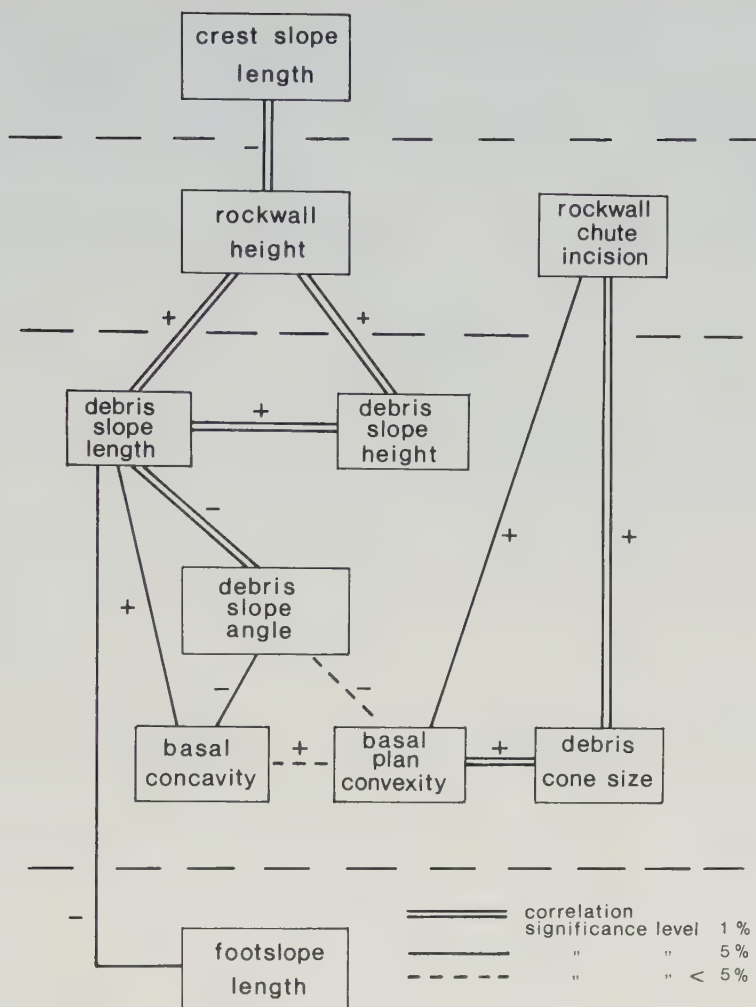


Fig. 4.26. Correlation bonds indicative of material transfer across debris slopes, affecting morphology and geometry between slope units in Nissunvagge. Based on measurements on Maps I and II (1:10 000 in original), using Kendall's rank correlation to establish bonds. (Values for tau between 0.42-0.67 at the 1 % level).

As shown by Map II, debris flows are very numerous along the valley sides of Nissunvagge. Their importance in debris slope development is especially marked on the eastern side below major chutes, where they are largely responsible for the observed relationships of Fig. 4.26 concerning basal concavity and plan convexity of slopes (cone formation).

The variable transporting capacity and range of debris flows of small vs. large size is a factor of interest not only with respect to total denudational effect, but also in terms of influence on slope form and particle size characteristics. Where small flows are predominant, the modification of rockfall talus toward 'alluvial talus' (Bones 1973, White 1980) with a basal

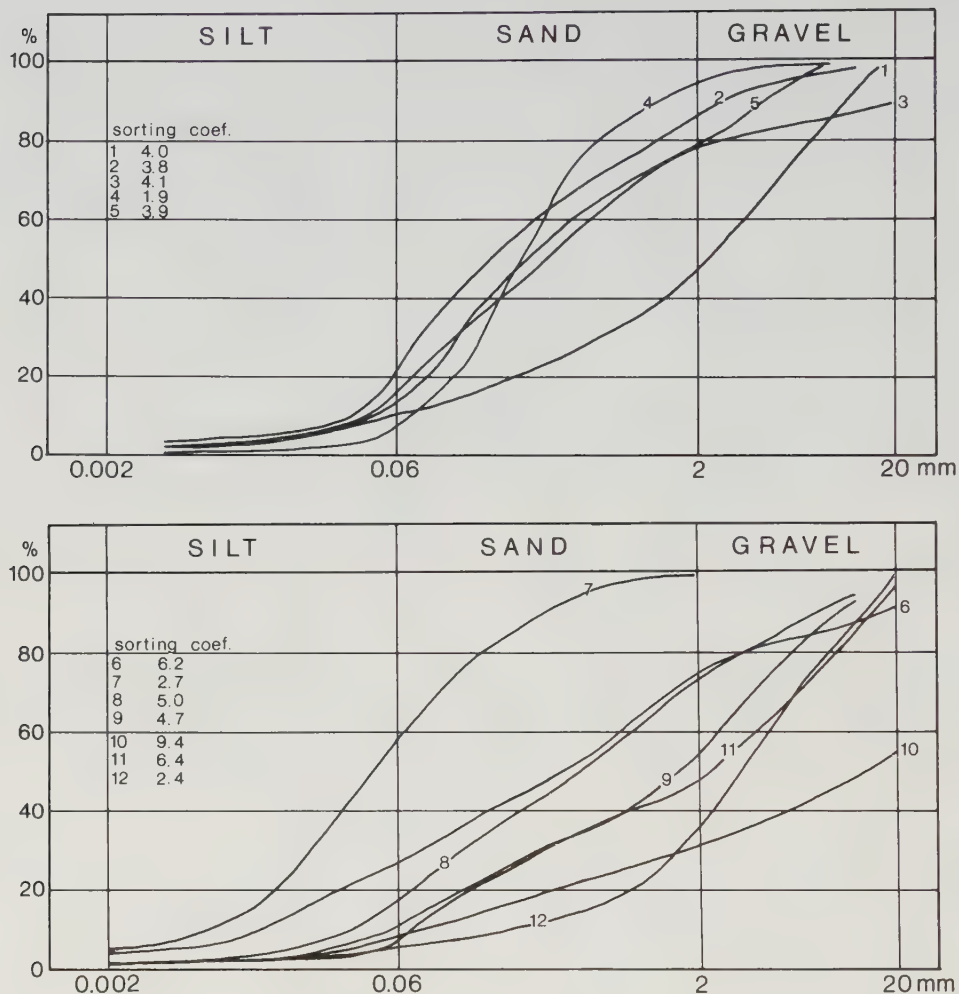


Fig. 4.27. Grain size distribution of debris flow material.

1. debris slide C5, zone of deposition
2. d. flow C42, core of frontal lobe
3. d. flow C38, lobe
4. d. flow C42, sandy outwash beyond frontal lobes
5. d. flow C42, levee
6. d. slide C5, scar
7. d. slide A13, scar
8. d. flow at Torneträsk station, source material along track
9. d. flow D23, source material in gully
10. d. flow C42, source material in gully
11. d. flow C42, source material in gully
12. d. flow at Tarfala, scar

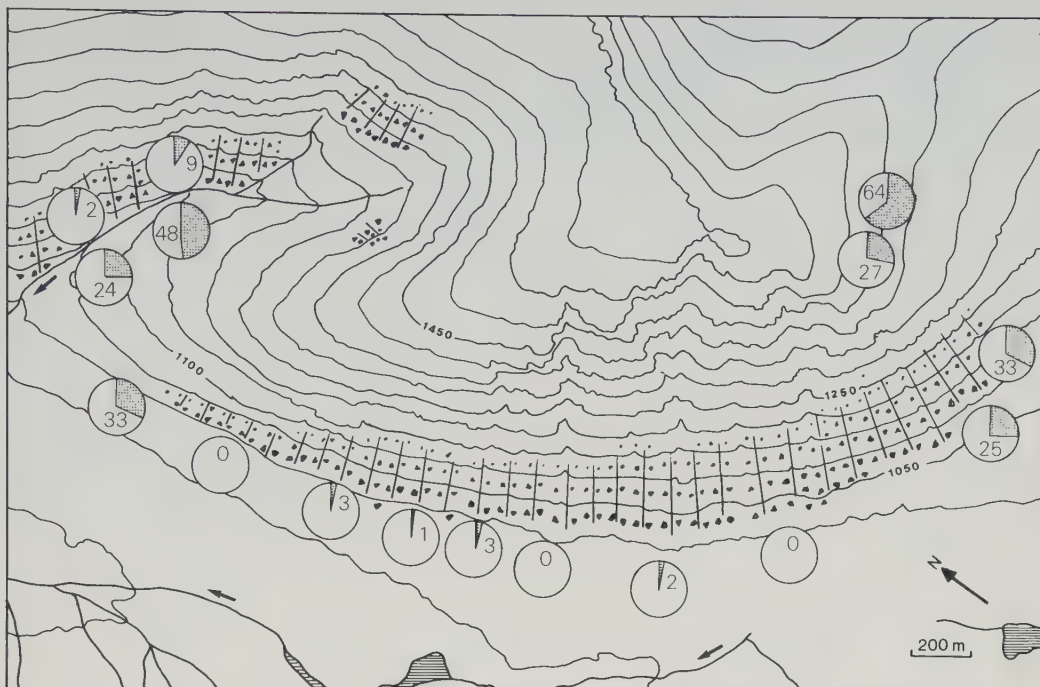


Fig. 4.28. Percentage of erratics found in debris flow lobes and in the blockfield at Mt. Nissuntjärro (sections B and C). Erratics are few on talus slopes, but common on till slopes, where no postglacial accretion of rockfall material has taken place.

concavity should be less marked compared to areas with more frequent large flows, such as Nissunvagge. Debris flows transporting mainly finer sizes will counteract the normal fall sorting of particles typical of talus slopes. They also introduce lateral variability in clast size and sorting (Bones 1973). Thus debris flow modification of rockfall talus probably varies in relation to the average magnitude of flows. Further data on the long-term geomorphological significance of debris flows in Nissunvagge will be discussed in section 4.4 and chapter 6.

D. Some material characteristics of debris flows

Material properties of debris flows are of significance to an understanding of their mechanism of movement and general capacity as a geomorphic agent. The general character of debris flow material is till-like (Fig. 4.27). It may be characterized as a frictional sediment with low amounts of cohesive particle fractions. With respect to coarse material, the uppermost parts of debris flows in Nissunvagge are occasionally clast-supported, although this may be a consequence of secondary winnowing of fines. The main part of the flows is typically matrix-supported.

The source material of debris flows in Nissunvagge has two different origins: 1) till with easterly erratics and 2) frost-weathered local debris, in the form of block fields and talus. This is seen in the variable amount of erratics found in debris flow deposits from different parts of the area (Fig.

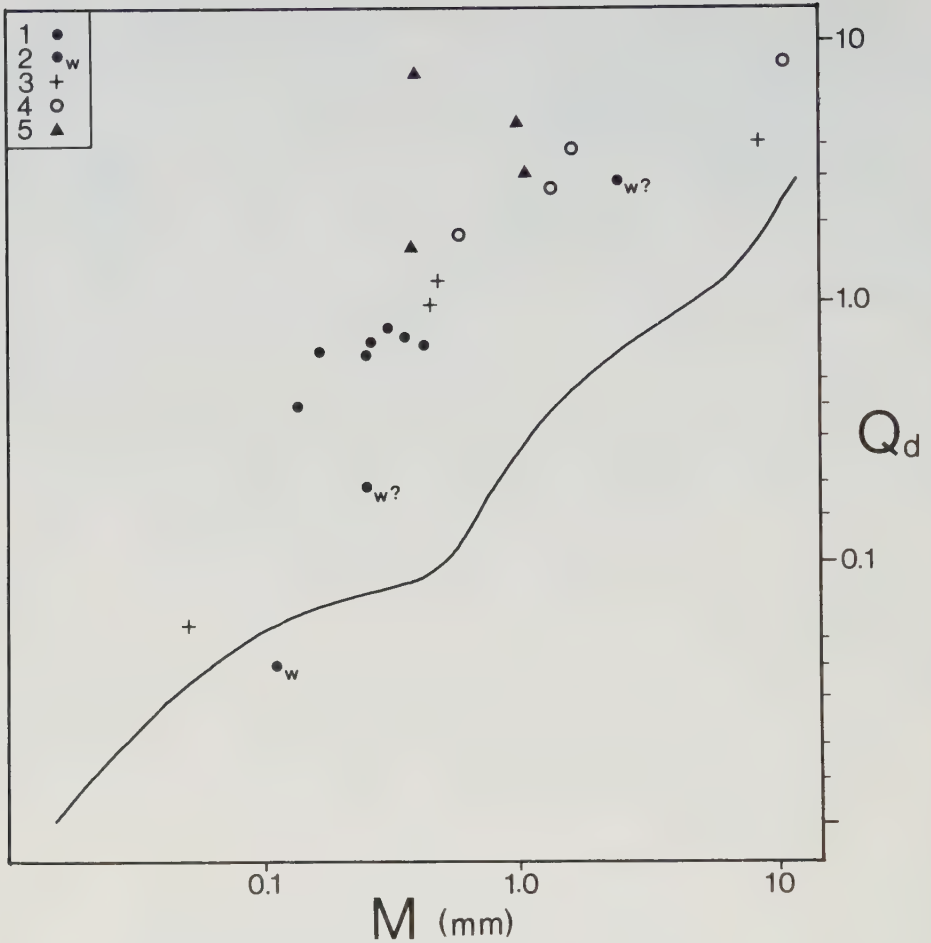


Fig. 4.29. Plot of median grain size ($M=d_{50}$) against quartile deviation ($Q_d = (d_{75} - d_{25})/2$) for debris flows and source materials in the study area. The diagonal line separates debris flow deposits from stream-flow deposits, according to Pe & Piper (1975). Key: 1. debris flow deposit 2. probably water-laid deposit associated with debris flow 3. scar of debris slide 4. till on slope with debris flows 5. till on stable slope.

4.28). Generally, flows with high amounts of erratics originated in till slopes or blockfields possibly derived from till, whereas flows with few erratics were always associated with slopes with talus overlying till or bedrock. Source material modification by the debris flow process is illustrated by Fig 4.29, where data for source areas and debris flow deposits are compared.

The majority of debris flows created in 1979 S of Abisko have a blocky surface. Sizes of clasts vary between 0.1-1.5 m roughly, showing the considerable geomorphic capacity of the process. Partly probably explained in terms of source material, different flows exhibit varying distributions of size classes (Fig. 4.30). Clast size plotted against debris lobe size gave positive correlations (max. 0.6, 5% sign.). Qualitative observations confirm that a distinction exists between small flows, less than 40-50 m

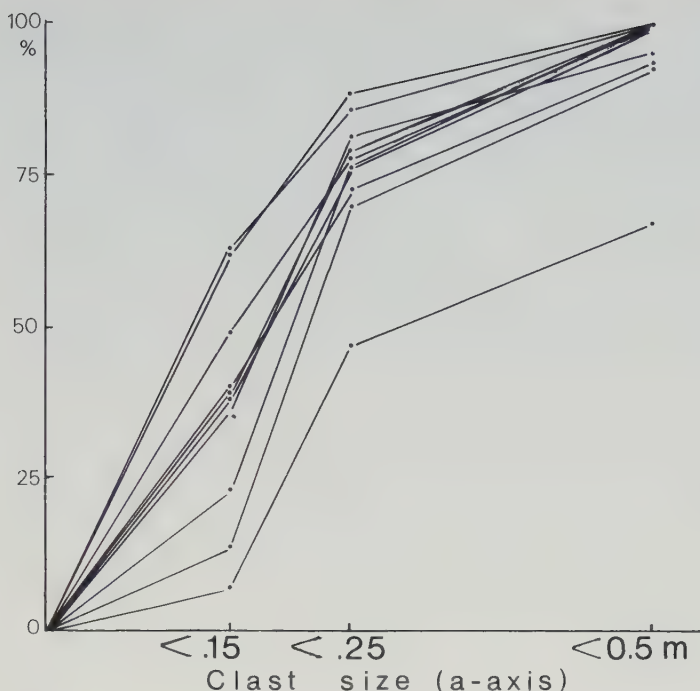


Fig. 4.30. Block size distribution on the surfaces of debris flows in Nissunvage. Based on transect counts across terminal lobes, systematically sampled.

long and flows of larger dimensions. The former usually consist of clasts of pebble- or stone-size (5-10 cm), in contrast to large flows with block-sized clasts (larger than 20 cm). The amount of finer material (clay to sand-sized fractions) on the surfaces of flows is also variable. No clear relationships were found between amount of exposed fine material and e.g. flow size or gradient.

A number of flows have been observed in sections. These display the expected internal structure with a core of a fine matrix supporting the surface blocks. Larger clasts occur also in the flow matrix, isolated or clustered. It is likely that distribution and orientation of these are indicative of flow viscosity, viscous flows having fairly uniform clast distribution while low-viscosity flows display tendency to graded bedding and horizontal or imbricated clast orientation (Bull 1977, p.237). The deeply incised channels at the apexes of debris flow cones afford some information on the stratigraphy of these features, showing that till underlies the local talus. However, fabric studies are needed to determine the genesis of the deposits in more detail. Possibly, a secondary redistribution of till material has taken place due to older debris flows (Fig. 4.31).

The orientation of surface clasts of debris flows is a function of the flow process. Particles tend to be orientated transverse to the flow movement in the lobes, often with a well developed imbricated structure. The particle orientation of levees is less uniform due to disturbances by blocks falling or sliding down the

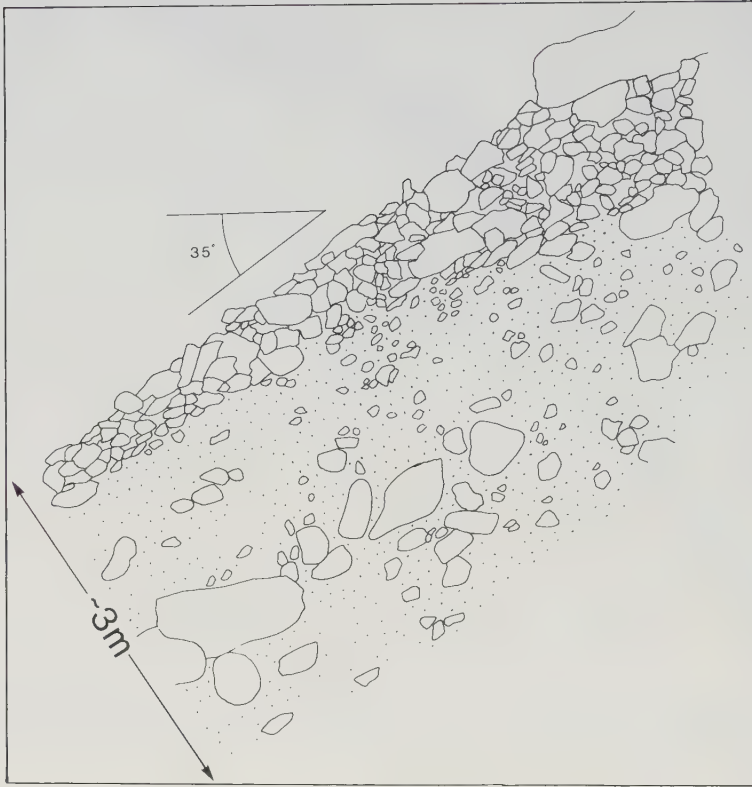


Fig. 4.31. Section through upper part of talus (debris flow channel A11). Drawing based on photograph. Beneath the upper coarse layer of talus material of 0.4-1 m, the section shows till or old debris flow material. The lower indicated horizon of coarse clasts may be related to older debris flows in the channel.

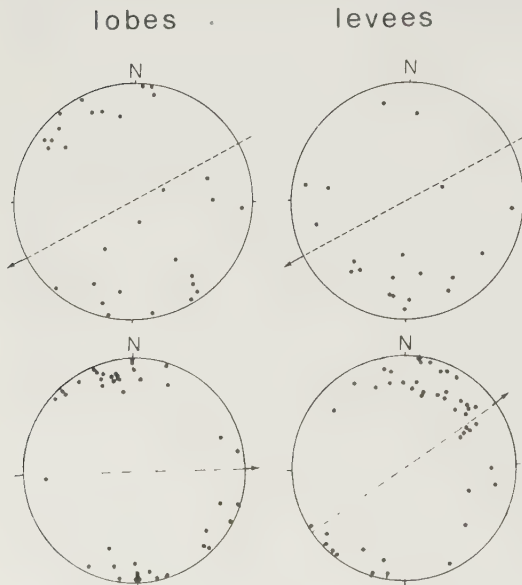


Fig. 4.32. Particle orientation of debris flows in Nissunvagge and Kärkevagge. Clasts on the levees show oblique orientation outwards from the flow, on lobes transverse orientation. Debris flow C33 Nissunvagge (upper diagrams) 7 and 18 Kärkevagge (cf. Nyberg & Brydsten 1977).

steep levee flanks. A tendency for oblique orientation with respect to the flow movement is visible in plots of surficial particle orientation of levees (Fig. 4.32). A preserved primary clast orientation may be a helpful diagnostic in separating old debris flow deposits from other features of similar morphology (cf. above). The fabric of the matrix material is probably more useful than surface clast orientation for this purpose, but is more arduous to obtain information on.

4.3.7 Slope form recovery after debris flow events

The recovery time of slope form following rare catastrophic events is a matter of significance to a proper assessment of the geomorphic importance of these events (Wolman & Gerson 1978).

Lichenometric data (section 4.4) show that old debris flow deposits may retain much of their original morphology for hundreds or even thousands of years. On the other hand, on some slopes debris flows have been strongly modified by other processes, e.g. snow avalanches (cf. Rapp 1975, Fig.8) or possibly by slow subsidence and resorting by frost action. Major changes likely to affect the initial morphology of debris flows are:

1. partial outwash of the fine matrix causing compaction and subsidence of the coarse debris of lobes or levees
2. subsidence and spreading of the flow deposits due to the influence on underlying material by the weight of the deposits
3. slow deformation of lobes and levees due to creep movements or frost action
4. infilling of the flow channel by material brought down from upslope or from the levees by wash or small-scale sliding
5. scouring of the flow channel by concentrated rainstorm runoff
6. scouring of levees and lobes and scouring or infilling of the flow channel by erosive snow avalanches

Evidently, these changes are partly opposite in their morphological effects. Little is known of the relative importance of the different processes. Kotarba (1981) notes a higher significance of avalanches and other mass movements on talus cones compared to fluvial activity in debris flow channels. Strömquist (1985) describes strong fluvial erosion of debris flow scars in Kärkevåg in connection with a heavy rainstorm in 1983.

To obtain information on the development of debris flow landforms following their initial formation, a number of sites will be periodically checked in the field, using different methods.

These observations have so far been of too short duration to allow any general conclusions on the rate of form adjustments at sites of debris flows in the area. The obtained fragmentary information from the years 1981-85 can be summarized as follows: Fine-grained debris flows, especially when deposited in thin levees and lobes with little erosion of the flow channel, are quickly modified through surface wash. Since the vegetation cover often remains intact beneath the flow, very few signs of the



Fig. 4.33. Small fine-grained debris flow at Lake Rissajaure, Kärkevagge, released at foot of rockwall and deposited on alluvial fan by the shore, without causing erosion on the grass-covered slope. Slope wash obliterated most signs of this flow in only a few years time. Photo 16.7.1978.

debris flow are left once the fine material has been washed down the slope. This may take place in only a few years time (Fig. 4.33), or possibly after a single heavy rainstorm, as described by Johnson & Rahn (1970 p.185).

Coarse debris flows undergo very small modifications during the first few years after formation. In a long time perspective, probably thousands of years, various processes may cause an obliteration or distorsion of their original shape. In some instances, debris flow channels become new permanent parts of the drainage network and may possibly even extend further upslope through the work of running water (cf. O'Loughlin 1972).

In conclusion, although periglacial slopes, as mentioned earlier, in some contexts are viewed as rapidly changing or evolving, the momentary changes caused by catastrophic geomorphic events appear to persist for long periods in nearly unmodified state, indicating a slow activity of 'normal' processes and thus large significance of the extreme events in slope development.

4.3.8 The effects of a rainstorm in July, 1982.

During the field work in Nissunvagge in the summer 1982 we witnessed the effects of a short, intense thunderstorm in the late afternoon of July 17. Although its impact on ourselves was considerable, the geomorphic impact was limited compared with the storm of 1979. As the rainfall amount was measured, a description of the event may help to further appreciate the extraordinary magnitude of the 1979 event.

Meteorological screen data from the preceding days and during the storm day are shown in Fig. 4.34. The weather had been fair during 15-16 July. Towards 1800 hours on the 17th, dark clouds were approaching Nissunvagge from S or SE and thunder was heard at a distance. At 1820 hours, rain commenced to fall in central Nissunvagge and a strong SE-wind replaced the previous weak N-NE-wind. The rain, accompanied by loud thunder, increased quickly in intensity. The heavy rainfall lasted about 45 minutes, by which time the wind waned. 14 mm of rain was recorded by the screen gauge, giving an intensity of about 19 mm/hour. The value is approximately equal to the limiting amounts for release of failures according to the rainfall intensity/duration relationship discussed by Caine (1980). As the rain was most heavy during the first 15 minutes, its intensity was briefly very high at its maximum.

Even larger amounts fell over the higher situated tributary valley (section B of the study). Rainfall distribution within the valley was studied during this event and during a less intense but longer lasting rainfall a few days afterwards. A number of raingauges (Pluvius) were installed in three cross profiles over the valley. The Pluvius gauge is subject to several errors. As the study had as its main objective to detect possible rainfall variations related to aspect within a small area (2 km²), the errors involved concerning absolute amounts were not considered. Each gauge was emptied into a graded cylinder at measurements, since no accurate calibration of the gauge readings had been made.

Rainfall distribution in the valley during the storm and the night of July 18 as indicated by the gauge network is shown in Fig. 4.35. These data can be interpreted in favour of a slightly higher precipitation on the northern valley side, which was the windward side during the storm, although the relationship is not clear for the lowermost profile. On July 23, small amounts of rain fell with northerly wind directions, thus opposite those of the 17th. In this case, differences between the valley sides were even less distinct.

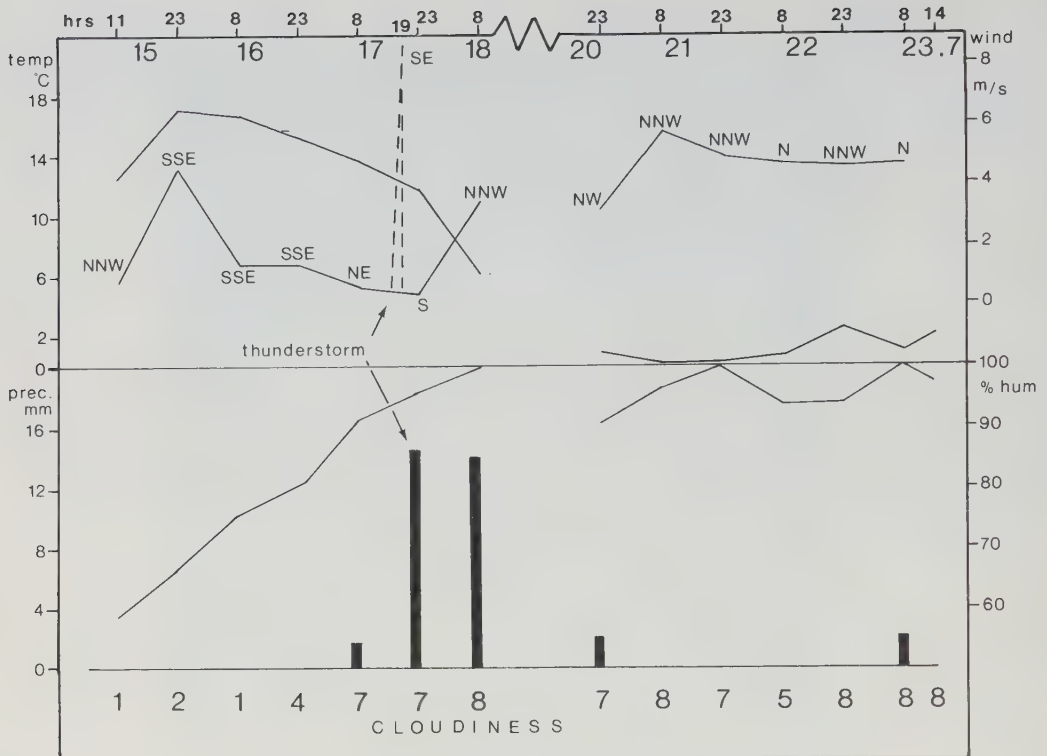


Fig. 4.34. Meteorological observations during a period in July 1982 with incidence of an intense thunderstorm in Nissunvagge. Cf. text and Fig. 4.35. Data from meteorological screen near mouth of tributary valley, section B.

The data on rainfall distribution are too restricted to allow any conclusions on the possible influence of slope aspect on rainfall intensity in the valley at the event or during the larger storm in 1979. Data by Strömquist & Rehn (1981) from the nearby Kärkevagge valley are aligned with the general view of non-uniform precipitation in mountain valleys, governed by the wind field as modified by local topography. With strong winds, lee-sides tend to obtain the highest amounts of rain (Geiger 1965, p.419).

Runoff from the valley was influenced by thick snowpacks filling the outlet or flanking the valley sides upstream. Overflow on snowbanks (Fig. 4.36) and a temporary fluctuation in discharge probably related to a snowbridge collapse were observed (cf. Woo & Sauriol 1980). Observations were also made of runoff in the old slide scars from 1979 on the grass-covered gelifluction slope in the northern part of section C, close to the mouth of the tributary valley. It was noticed that the scars served as new drainage lines concentrating runoff. In the upper parts of the scars, small mudflows were set in motion, coming from the steep exposed sides of till in the scars. As the mudflows became channelled some few meters downslope, they mixed with further water and turned into muddy rill flow. This mobilization of silty

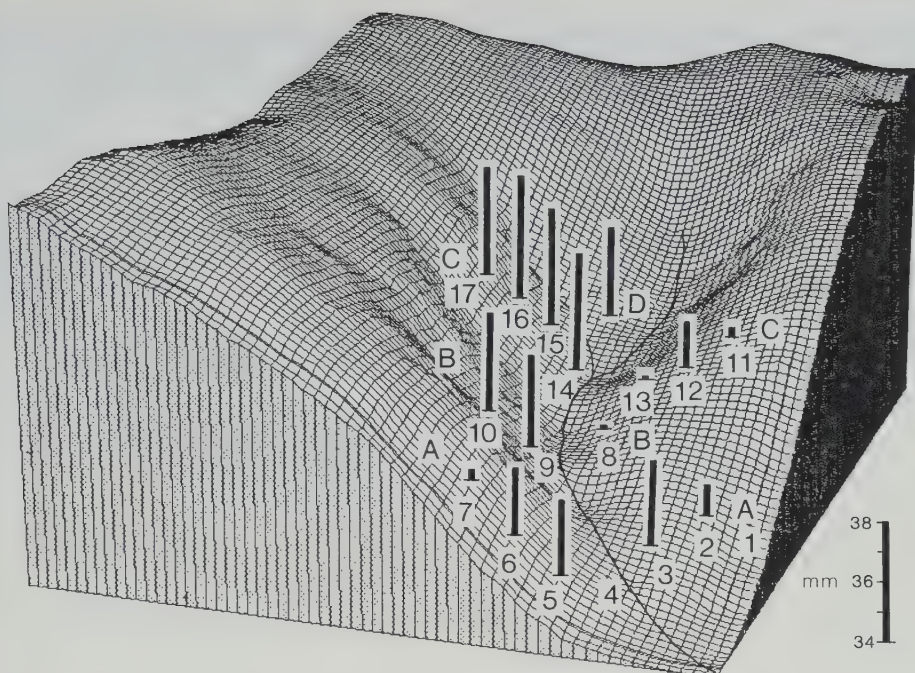


Fig. 4.35. Rainfall distribution in the tributary valley (section B), Nissunvagge, during the thunderstorm event July 17, 1982. Measurement points were along three transverse profiles, A-C, used also for air and ground temperature measurements. Note the scale of the histograms, showing precipitation amounts in excess of 34 mm, to highlight differences between the valley sides. The southfacing side received slightly higher amounts during this event, with the rainstorm movement being from south to north.

material and its outwash from the slide scars by running water resulted in thin sheet deposits on the terrace beneath the slope. Only three new debris flows were released, two on the south side of section B and one in section D (see Map I). In all three cases, the small flows or slides had been superimposed on snow banks (Fig. 4.37).

Discussion

The geomorphic effects of the storm were mainly the erosion and transfer of material by running water. Mass movements were of much less significance during this event than for the 1979 event, implying a critical threshold of rainstorms, shifting or broadening their geomorphic impact from fluvial to mass movement processes. This threshold could thus be somewhere around rainfall intensities of 20 mm/hour in this area. Of course the activity of different processes is also related to slope angle and other factors, which complicate the issue. However, it is clear that the 1979 storm was an effective 'formative' agent, whereas the 1982 storm mainly caused some minor denudation with small effects on landforms (cf. Wolman & Gerson 1978). The available observations have not permitted any estimate of the amount of denudation accomplished by the storm of 1982 to be made.



It should also be noted that the washing-out of weathered debris in slope depressions by a major storm event, as in 1979 in Nissunvagge, may influence the effectiveness of following storms, as little transportable material is available. Observations probably showing this effect have been made e.g. at Disko, W. Greenland, where a storm of 62 mm on August 15, 1984, causing numerous debris flows near Godhavn, was followed after 7 days by an event with 31 mm rainfall, which had no or small geomorphic effects (O. Humlum, pers. comm.). See also discussion in Innes (1983, p.585) concerning progressive weathering.

4.3.9 Summary and conclusions

This section has exemplified the geomorphological effects of debris flows in an arctic/alpine environment by the description of the most recent and largest known case of these mass movements in the Swedish mountain area. Although no direct observations were made of the debris flows in Nissunvagge at the time of the event, morphological features of the flows, i.e. the landform record, seem to agree with previous reports of debris flow motion elsewhere. The occurrence of several pulses within a single flow seems to be common, for larger flows. The long downslope range of some flows across gentle footslopes may seem remarkable in view of the generally very coarse debris involved, but has been given possible theoretical explanations (cf. Rodine & Johnson 1976).

The following conclusions can be drawn:

1. Primarily steep slopes ($>30^{\circ}$) were affected, giving a marked uneven occurrence of debris flows in small asymmetric valleys.
2. Concentration of runoff to concave spots in the terrain is considered primarily responsible for the localization of the release areas. Consequently many debris flows followed old drainage lines or chutes along the slopes.
3. The most important factor releasing the mass movements was most likely high pore water pressures in the thin till or debris mantle on the slopes, following the intense rainfall.

Fig. 4.36. Torrential runoff following intense thunderstorm rainfall over central Nissunvagge on July 17, 1982. Before the event, the discharge of the stream from the v-shaped valley in the background (section B) took place under the snow-drifts filling the valley entrance. For a short period of peak flow during the storm, runoff occurred on top of the snowbanks. The water was brown-coloured of debris, eroded along the stream banks and in the channel upstream in the catchment. Photo Lars Barring 17.7.1982.

Fig. 4.37. Small debris slide caused by the rainstorm on July 17, 1982, in the tributary valley, southern side. The failed masses were deposited on top of a remaining snowbank along the stream channel. After melting, the material will be temporarily stored as a debris cone along the channel, until it is eroded and transported further downstream by future high water events. Photo 18.7.1982.

4. Some chutes periodically serve as channels for debris flows, which deposit their material as cones of debris further down-slope. Although other processes also influence the cones, they are dominated by debris flows of long recurrence interval in their morphology.
5. As debris flow cones are numerous in Nissunvagge, they show that a) flows have been important before 1979, probably during the whole postglacial period, and b) debris flows often only lead to a local modification of slopes within the catchment.
6. On steep slopes undercut by a stream, the material transported by debris flows constitutes input to the stream channel system and is more quickly removed, giving a direct influence on denudation calculated on a catchment basis.

4.4 Frequency of debris flows in northern Swedish Lapland

4.4.1 Previous information on debris flow frequency

Debris flows occur irregularly, often with long time intervals for a specific site or within a limited area. As noted by Johnson (1970), who used laboratory simulations to study debris flows, 'the student of debris flow might be an old man before he had the opportunity to analyze a natural debris flow' (Op.cit.p.497). Only in some environments, such as the Japanese Alps do they recur as frequently as every year on certain sites (Okuda et al. 1980). Available data suggest that the frequency varies for events of different size. It also appears that flows of different magnitude may have contrasting frequencies in different areas, some areas being characterized by a preponderance of small flows and others by larger ones (Innes 1985b).

Knowledge of debris flow frequency (occurrence in time) is based on direct observations of flows as reported in the literature, and on indirect evidence drawn from the morphology and vegetation of older flows. Available precipitation records from weather stations near affected areas may give some information on the average expected recurrence interval of extreme events triggering mass movements (cf. Rapp & Nyberg 1981b). However, the local nature of thunderstorms limits the usefulness of data from valley stations often tens of kilometers distant from the high mountain areas experiencing the induced geomorphic activity. The extrapolation is further complicated by the need to consider the rainfall amounts preceding the triggering event (Strömquist 1976).

Rapp (1960b) discusses threshold values of precipitation necessary to initiate mass movements in northern Scandinavia, and suggests 50 mm in 24 hours as a geomorphologically active rainfall. The event of October 1959 with 107 mm/24 hours is considered an extremely rare, centennial or even millennial, maximum (Op.cit.p.185). Millennial was apparently an exaggerated estimate, as the Nissunvagge case in 1979 shows. In the long term, repeated processes of this type are nevertheless assigned an important role in the slope evolution in this area, although probably not comparable to their significance in areas of more frequent torrential rains and higher relief, such as the Alps (Rapp

1962, p.297). Strömquist (1976), presenting the geomorphic impact of this event at Andöya, suggests a recurrence interval for this type of widespread slope instability due to rainfall of about 40-50 years. Rapp (1974,1975) also describes the effects of a rainstorm on July 6, 1972 at Tarfala, where numerous debris flows were released. From the time the glaciological research station started its activities in the valley in 1947, no large debris flows have been reported apart from the 1972 flows. As noted by Rapp (1975) and later by Rapp & Nyberg (1981a), older generations of debris flows occur in the valley (see below). A further modern case of debris flows occurred at Tarradalen west of Kvikkjokk on the same day as the case at Tarfala (Leif Engh, personal communication). Rapp & Strömquist (1976) summarize information on several cases of rainfall-triggered mass movements in the Scandinavian mountains.

As mentioned previously, debris flows have been reported from nearby parts of northern Scandinavia, by Rudberg (1950) and Zenzén (1926) with general comments on their frequency. A case from a more southerly position in the Swedish mountains is that mentioned by Ångeby (1947, p.102), referring to a torrential rain in Jämtland causing slope erosion in 1938. The geomorphological maps from the Swedish mountain region show a widespread distribution of debris flows. However, the maps give no information about the age of the forms.

In this study, further data on debris flow frequency have been sought from datings of vegetation on deposits or along tracks.

4.4.2 Assessment of debris flow frequency from morphology and lichenometric data.

Cross-cutting of levees and overlapping of lobes form a morphological basis from which some inferences can be drawn concerning process frequency. The complication in this respect, already mentioned, is the fact that several successive surges may occur during the same flow event, and therefore further data should be utilized to positively distinguish debris flows of different age on a particular site. The lichen cover on the deposits has proven to offer a possible means of separating debris flows with clearly different ages, with the size of individual lichens as an indication of age (time elapsed since colonization by the lichens). The application of the lichenometric method and problems inherent in this technique were dealt with in the section on methods (3.5). The previous work by Karlén (1973,1975) within the study area has shown the usefulness of the technique. Innes (1983) has applied lichenometry to dating of debris flow deposits in Scotland. Four localities were used for lichenometric studies concerning debris flow frequency: the Nissunvagge and Alesättn areas (bedrock: amphibolite) with the latest debris flow phase in 1979, the Tarfala valley (schist, amphibolite) with the latest phase in 1972, and Mt. Ertektjåkko (schist) in the southernmost part of the Norrbotten mountains, with unknown age of the latest debris flows, but with numerous flows within a short distance. Tentatively, Karlén's (1975) results of no marked variations in growth rate for different bedrock types are accepted in applying his growth curve to data from all four sites. The results of the measurements, shown in Table 4.4, represent a preliminary step towards the development of a debris flow chronology in these areas, since the number of old flows available in the terrain is very large. Only a limited number of them could be included in

Table 4.4 Lichen diameters (subgenus *Rhizocarpon*) on debris flow and other mass movement deposits, northern Lapland. Estimated ages derived from the growth rate curve by Karlén (1973).

locality and position	single largest lichen diameter mm	mean of 5 largest mm	estimated age, years before 1980
NISSUNVAGGE (5 sites)			
section A, d.flow near A4	170	163	2200
" " A12 a	185	163	2600
" " b	158	133	1900
C42, lobe 1 (Fig.4.39)	183	168	2600
" 2	95	90	415
" 3	99	91	450
" 4	50	46	160
" 5	90	80	365
" 6	96	90	415
" 7	85	79	335
" 8	88	85	355
" 9	84	82	335
" 10	76	73	250
" 11	135	123	1300
" 12	140	135	1500
talus adjacent to C42	220	-	3400
section D d.flow D2 lobe 1	110	102	700
" 2	106	92	600
" near D23 lobe	107	104	600
" leve/gully	100	94	450
ALESÄTNO (1 site)			
Pakkapahuktjåkka S.slope, d.flow lobe (Fig.4.9)	55	53	180
TARFALA (4 sites)			
lower valley, E side, d.flow no.2 from the south, lobe a	83	81	330
" " b	40	37	120
d.flow no.3, lobe a	180	169	2500
" " b	80	75	310
" " c	140	124	1500
d.flow no.4, lobe a	158	145	1900
" " b	71	68	250
" " c	80	75	310
talus between flows 3 and 4	181	156	2500
upper valley, E side, N of 1972 flows E of Res.Stn			
d.flow lobe a	98	93	440
" b	114	105	750
" c	144	131	1600
blocks on slope adja- cent to d.flow above	220	203	3400
ERTEKTJÄKKO (5 sites)			
N. slope, d.flow no 3 from the east	66	62	230
" no 5	55	54	180
" no 10	60	56	205
" no 11a	60	57	205
" no 11b	56	54	180
" no 12	45	44	140
MT.VIDJA rockslide	350	328	6800

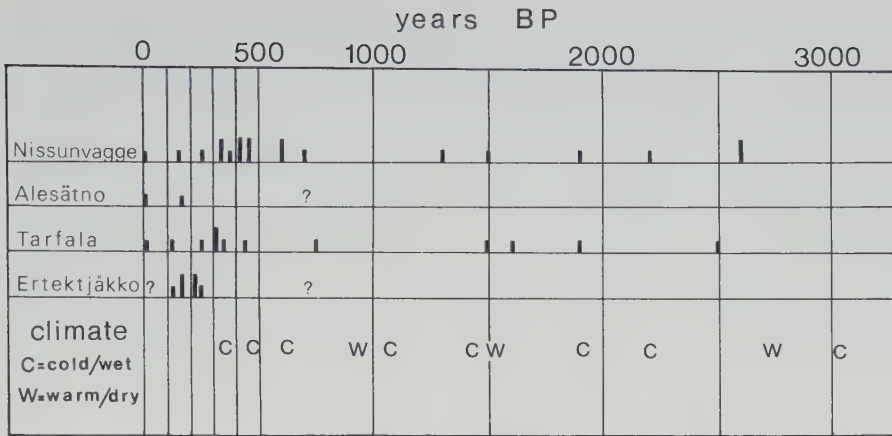
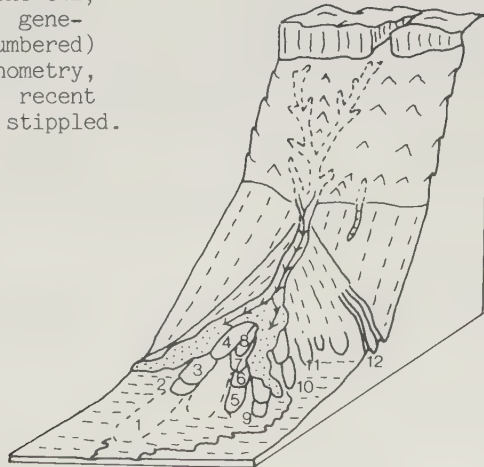


Fig. 4.38. Tentative datings of debris flow deposits by adoption of the lichen growth rate curve for northern Lappland by Karlén (1973). The histograms represent one or two dated cases each. The record is incomplete, especially regarding Alesättno and Ertektjåkko, where only a small number of the total available number of flows have been dated. Nearly contemporaneous events during the last 500 years may represent identical events (release of flows as 'cells' of rainfall impact), due to uncertainties in the lichenometric dating technique. Likely variations in climate have been indicated, based on Karlén (1982). Many debris flow events appear to correspond to cold periods, but the datings are too uncertain at present to allow firm conclusions to be drawn.

Fig. 4.39. Sketch of debris flow cone C42, Nissunvagge, showing different generations of levees and lobes (numbered) which have been dated by lichenometry, cf. Table 4.4. Fresh debris of recent debris flow event (1979) shown stippled.



the measurements, embracing 37 debris flow lobes or levees, distinguished on morphological grounds and occurring at 15 different sites of debris flows. Although a tentative dating of the flows has been made, this should be seen as a first attempt to compare the periods of debris flow events with climatic data and to allow an illustration of the relative spacing of events to be made.

A. Interpretation and discussion

The data indicate a number of debris flow events of different ages, as the lichen sizes on the various flows can be grouped into classes separated by intervals. Tentatively, the datings may be structured into between 6-10 separate events for Nissunvagge and 7-8 for Tarfala during a time period of 2500-2600 years (Table 4.4, Fig.4.38). The lower numbers are the more conservative taking into account not only the largest lichen but also the mean of the 5 largest individuals. The size variation within some classes is large, representing probably several hundred years of age. In these cases it is difficult to judge whether the class consists of several debris flow events closely spaced in time, or if a single event may show this large spread in lichen sizes, due to environmental factors. Thus the present data do not allow a precise determination of the number of old debris flow events at the investigated localities.

There is a closer spacing of events during the last 500 years than earlier. This could indicate an increased frequency of debris flows, but may also reflect difficulties in detecting very old debris flows as morphological units in the terrain, e.g. if they have become subsided and restructured into patterned ground, or simply buried beneath more recent debris flows. It should also be pointed out that the dates of old flows are more uncertain than for the more recent ones, as the growth curve for lichens by Karlén (1973) is based on simple extrapolation for large lichens, and as discussed, may be less reliable in this part due to interspecies growth variations (cf. Innes 1982). The lack of very old flows was noted also by Innes (1983) working in Scotland. Most flows had occurred within the last 500 years. A decline in the vegetation cover related to landuse changes was considered important, rather than climatic variations, in explaining the increased frequency of debris flows. In Swedish Lapland, this hypothesis is less probable, due to the low impact of human activities.

Grove (1972), using historic data, notes a period of increased frequency of landslides between about 1650 to 1760, with a peak in the last decennium of the 17th century. Both Nissunvagge and Tarfala had debris flow events during this interval, according to the lichenometric data.

Karlén (1982) has found indications of a periodicity in climatic variations of about 200-300 years in Holocene time, based i.a. on studies of glacier fluctuations. This periodicity is fairly similar to the average recurrence interval of debris flows during the period 500-2500 years BP, if we assume that the maximum sizes of lichens represent separate events (Nissunvagge 7 events= 286 years, Tarfala 5 events= 400 years). Only a few direct correlations between dated debris flows and cold/wet periods are evident in Fig. 4.38. In view of the present uncertainty of the datings both of debris flows and climatic periods, a good coincidence is

hardly to be expected. The datings of overlapping lobes of cone C42, Nissunvagge (Table 4.4, Fig. 4.39) suggest uncertainties in the technique up to 80 years, assuming the lobes resulted from successive surges of isolated events.

A few measurements on substrates other than debris flows are included in Table 4.4. Three datings of talus or block slopes where the debris flows had occurred gave preliminary ages of 2500, 3400 and 3400 years. The taluses had low recent activity and their lower parts, where the measurements were made, appeared stable with well developed lichen cover. The rockslide at Mt. Vidja (cf. Rapp 1960b, p.119) was tentatively dated at a minimum age of 6800 years, although a larger sample of measurements from other parts of this very large deposit is needed to support this date. However, the measurements suggest an early postglacial age.

B. Concluding remarks on the lichenometric datings

The dated ages assigned to debris flow events in this study may be subject to revision later on, in the light of additional information. For the present purpose of estimating debris flow frequency, the actual number of discrete events and their relative spacing in time are the main interests, rather than an absolute time scale. Although the lichenometric technique of dating is only approximate for these old features, the data indicate that the debris flows investigated in the study are not of late glacial or early postglacial age, further that the taluses that they dissect have continued to develop since early postglacial time, as they show younger ages than the rockslide at Mt. Vidja, where deposition was momentary and the blocks were left open to lichen colonization and growth without disturbances of subsequent deposition of material.

4.4.3 Tree-ring dating

The long debris flows at Alesättno and near Torneträsk station were the only ones of the 1979 event that reached into forested areas (Figs. 4.7, 4.40, 4.41), thus offering opportunities to estimate flow frequency from the age of the tree population. The largest trees felled by the 1979 debris flows at Alesättno were dated to between 65-80 years of age. This gives a broad indication that debris flows of comparable size to those of 1979 have not occurred at this site during the present century.

4.4.4 Summary and conclusions

Previous field observations and consideration of meteorological data have indicated fairly low frequency of debris flows in northern Scandinavia, with several decades or hundreds of years between major events. The attempts in the present study at dating old flows from their lichen cover or from their effects on birch forest along the tracks, have yielded similar results, with a possible indication of increased frequency during the last 500 years compared to the period 500-2500 years BP. No flows of older age were encountered. The lichenometric dating method is fraught with several inaccuracies that may question the absolute ages obtained or the separation of nearly contemporaneous events. However, in northern Lapland it nevertheless seems to be a useful method for the study of mass movement frequency, with a larger range in time and a more general application than dendrochronology or other botanical methods (Sauchyn et al.1983),



Fig. 4.40. Debris flow near Torneträsk station, most likely released on June 23, 1979. A debris fan of 5° inclination was formed downslope of a u-shaped debris flow channel with levees. The flow felled several mature mountain birches and almost reached the railway line (cf. Fig. 4.41). Photo July 1980.

owing to the predominant occurrence of mass movements in the tree-less high mountain area. For the present, the results based on lichenometry must be viewed with caution, until more research on the reliability of the method has emerged.

4.5 Regional distribution of debris flows

4.5.1 Available data

Apart from the few cases of debris flows described in the literature, the available information on the regional distribution is provided by the geomorphological maps of the area (Melander 1975a,b, 1976a,b, 1977a,b, 1982, Ulfstedt 1977, 1979, 1980, 1982, Hoppe & Melander 1979). Although they present the distribution in a generalised manner, thus limiting the usefulness of the data for numerical analysis, they afford some measure of the relative abundance of debris flows in different parts of the area. However, the maps are based on air photo interpretation and thus most likely under-represent old flows not easily identified on the

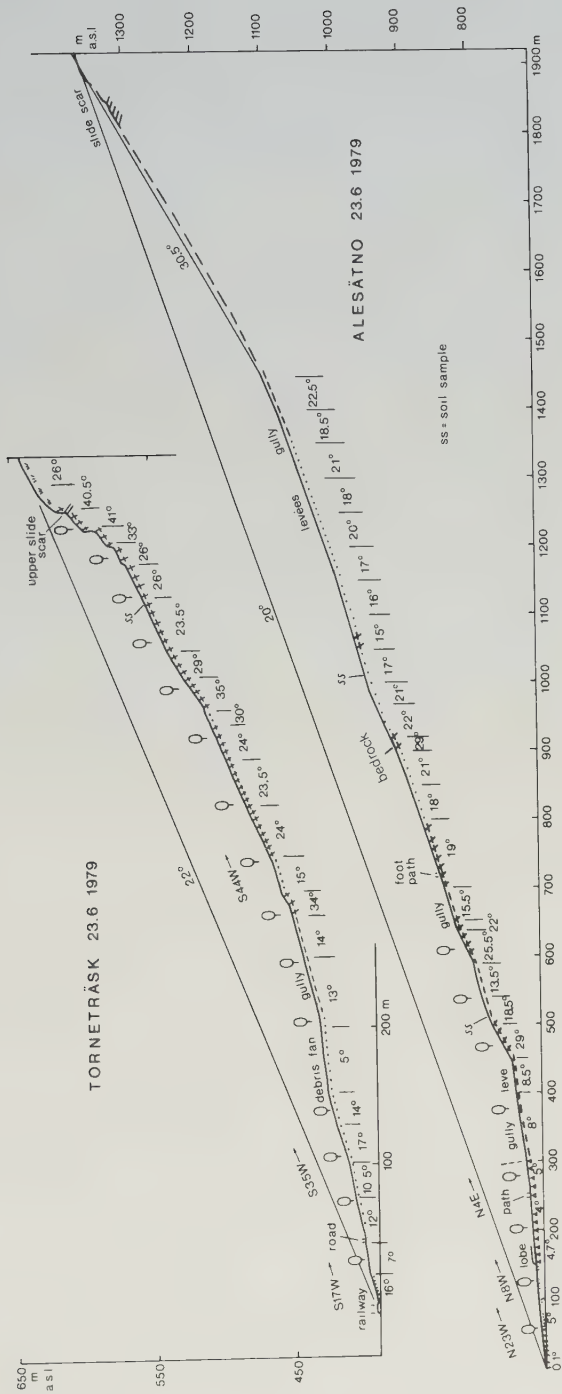


Fig. 4.41. Debris flows of 1979 with effects on mature mountain birch forest. Dendrochronological datings of the largest trees felled by the flow at Alesättn indicate at least 80 years of stable growth conditions and hence, absence of previous large debris flows. Note the influence on hiking path and railway embankment by the flows. Cf. Figs. 4.7 and 4.40.

photos. Furthermore, and possibly more problematic for the complete picture of debris flow distribution, there is a selective representation of dominant forms where several different types occur. Thus debris flows often dissect avalanche boulder tongues, e.g. at Mt. Ertektjåkko, S. Norrbotten, but only the boulder tongues are shown by the map. Less serious is probably the occasional confusion of debris flows with slush avalanche tracks and deposits.

4.5.2 Distributional patterns

An overview of the distribution of debris flows within the study area has been compiled from the geomorphological maps (Map III). It is evident from the map that debris flows are strongly concentrated to areas of high relief, especially the amphibolite mountains from the Tuoptetjåkka massif north of lake Torneträsk to the Sarek area. There is a remarkable absence of debris flows in the southern part of the Norrbotten mountains, which may be attributed to the generally lower and less steep mountains compared to the more northerly areas. However, some massifs of considerable relief exist, e.g. the Jeknafo, Saolo, Tjiddjak and Kustar mountains, where the lack of debris flows is surprising. Probably a more detailed search for older less distinct flows would fill in the gaps in the distribution for this part of the region.

The comparatively low frequency of debris flows in the Kebnekaise area according to the geomorphological maps may be a corollary of the high frequency of avalanche boulder tongues there, taking preference over debris flows in the representation on the maps. In the Sarek area, where avalanche boulder tongues are somewhat less common than in Kebnekaise (Hoppe 1983), debris flows are more numerous on the maps.

Discussion

Debris flows seem to be initiated in variable numbers in relation to the type of rainfall inducing the mass movements, either in the form of isolated 'spots', limited areas or 'cells', or finally extensive areas or 'regions' of rainfall geomorphic impact (Rapp & Nyberg 1981b, Rapp 1985). One may speculate on the significance of this spatial occurrence pattern with respect to the cumulative distributional pattern of debris flows. According to this hypothesis, local thunderstorms such as the one over the Nissunvagge area in 1979 may give rise to anomalies in the overall distribution of debris flows that cannot be accounted for by differences in relief, soils etc. Naturally, the likelihood of such a pattern being recognizable is dependent on the frequency and spatial spreading of debris flow-triggering rainfall events in postglacial time.

On Map III, a large spot (more than 6 cases), or several adjacent large spots, may correspond to 'cells' of simultaneous release. E.g. the Nissunvagge 1979 event is represented by 6 large spots and two smaller ones. The majority of smaller spots correspond to isolated events or widespread scattered release over large areas by frontal rainfall or rapid snowmelt. Assuming an average of 4 large adjacent spots, or 1 if isolated, to represent a major rainfall event causing debris flows, the number of cells in the area of Map III is about 30-35. During 9 000 years of postglacial

time, a recurrence interval of events spread over the area of 250-300 years is needed to cause this pattern. However, as the evidence from Nissunvagge and other localities shows, the same area can be struck by debris flow events repeatedly, although the latest event at Nissunvagge was sufficiently large by its own to cause the distributional pattern for this area of the map. Therefore it seems reasonable to believe that many more than 30-35 events have occurred in postglacial time, even if renewed occurrence of mass movements on a particular site must await the attainment of threshold-related levels of debris production by weathering. The reason why their geomorphic impact is unevenly spread in the region may be due to a topographic influence on rainfall, concentrating the extreme events to the high relief areas, or if heavy rainfalls are common also elsewhere, imply that the less steep slopes in these areas are more stable with respect to rapid mass movements initiated by high pore water pressures. As indicated by the present study of debris flow occurrence in the Nissunvagge area, slope angle appears to condition the distribution on a local scale, in conformity with the distribution of snow avalanches. However, the clear relationship to wind regime and snow drifting observed for the aspect of avalanches does not apply to debris flows, the orientation of which ought to be more variable on a regional scale (cf. however Åkerman 1984). No detailed analysis has been carried out of debris flow aspect regionally in the present study.

4.6 Debris flows as natural hazards

4.6.1 The concept of natural hazards in mountain areas

A natural hazard can be any of a number of natural events, usually rapid, which pose a threat to human life or property. Along with earthquakes, volcanic eruptions, rock slides and floods, avalanches and debris flows belong to a class of hazards sometimes referred to as 'geomorphic' or 'geological' (Bolt et al. 1975). Phenomena of these types are often also referred to under the heading of 'natural catastrophes' (Scheidegger 1975). Research on natural hazards has gained considerable momentum in recent years, often with the aim to produce hazard maps displaying hazard zones (e.g. Kienholz et al. 1984). In mountain areas, the steep terrain makes various types of mass movements particularly dangerous to people and animals. Snow avalanches and large rockslides ('rock avalanches') have attracted the most attention, although also debris flows have been noted as mountain hazards (e.g. Grove 1972, Nasmith & Mercer 1979, Okuda et al. 1980, Hollingsworth & Kovacs 1981, Sauchyn et al. 1983). A general account of the hazard associated with debris flows is given by Costa (1984).

The limited treatment of the subject here is mainly focussed on the terrain and climatic conditions characteristic of the Swedish mountains with respect to hazard. As far as is known, no fatal accidents have occurred due to debris flows in Sweden. This is most likely a consequence of the sparsely inhabited nature of these mountains, as well as moderate although increasing tourism,

compared with other mountain areas with frequent accidents, e.g. the Japanese Alps, where about 90 people are killed each year by debris flows (Takahashi 1981). However, with increasing development of the Swedish mountains, debris flows are among the natural events that can be expected to cause problems. The new road from Kiruna to Narvik was blocked for traffic by a debris slide/flow on May 17, 1984, below the steep slope of Mt. Njulla, within a section of the road that is furthermore subject to snow avalanche danger. No doubt, the hazard by debris flows must be recognized as real, considering the reach of the 1979 flows in Nissunvagge and at Alesättno, affecting gentle, grass-covered slopes suitable as camping ground (Fig. 4.7). In the following, a brief outline of potentially dangerous terrain sites and the weather situations most likely associated with the hazard of debris flows is given.

4.6.2 Hazardous terrain sites for debris flows

Any steep slope ($>20^\circ$ incl.) with a cover of till or regolith (weathered rock debris) with included fine material can be affected by debris flows. Some locations are more likely as should be clear from the previously given description of site characteristics of debris flows (ch.4), in particular parts of a slope where water can be concentrated in the upper ground layers. As steep slopes are intuitively recognized as dangerous by most people, they will not be emphasized further here. It seems more justified to point to the hazard on slopes of more gentle inclination, where no danger is generally perceived. Concerning debris flows, two forms of low angle slopes are hazardous, not as release areas, but as depositional zones. These are

1) the bottoms of narrow valleys issuing from a mountain massif, sometimes incised in alluvial fans. These locations are suitable as camping sites only if level ground occurs in the valley bottom, e.g. as old revegetated stream beds beside the present one. Such sites may be flooded or overrun by valley-confined debris flows issuing from the headwater area in connection with heavy rainstorms (cf. Fig. 4.13).

2) gentle footslopes (about 10°) in front of often steep rock walls and talus. Such slopes may appear as safe sites for camping, but debris flows may extend for several hundreds of meters across them. Occasionally it is possible to detect the potential hazard of these slopes from old debris flow deposits. Ordinary stone and rockfalls are usually no danger on footslopes, and the low recurrence interval of debris flows makes them safe except during extremely heavy rainfall events. Cf. Figs. 4.7, 4.41, 4.42.

The approach of a debris flow is accompanied by loud rumbling sounds (Blackwelder 1928) or ground vibrations (Okuda et al. 1980), which may provide some short-term warning.

4.6.3 Meteorological conditions associated with increased hazard

As mentioned before, debris flows are initiated by high pore water pressures following extreme rainfall, or rapid snowmelt. Also antecedent weather conditions and soil moisture before a new input of water are important, as the slopes may be more or less close to a threshold stability condition. It is therefore difficult to state in a general way any particular rainfall amounts or intensities that are liable to cause the release of debris flows.



Fig. 4.42. Eastern part of Mt. Nissuntjärro (section D) with debris flows from 1979. Unusually long flows occurred on the till slope to the east of the rockwall and talus. One flow extended for several hundred m on a gentle gelifluction slope (arrow), showing the potential hazard of debris flows to hikers. Air photo N.Å.Andersson 25.7.1979.

However, it appears that the hazard is increased if a rainfall is short and intense, rather than of prolonged duration and lower intensity. Hence thunderstorm downpours are probably the meteorological conditions most critical with respect to debris flow hazard, an assumption supported by the cases of Nissunvagge in 1979 and Tarfala in 1972, as well as by many other cases of debris flows elsewhere.

4.7 Summary and conclusions

Debris flows are widespread in the high mountains of northern Swedish Lapland. The most common terrain position is as hill-slope flows, usually dissecting talus. Larger valley-confined flows are more rare, or leave their deposits e.g. on large alluvial fans where they comparatively rapidly are reworked by other processes (flash floods, slush avalanches).

Debris flows are initiated by high pore water pressures in loose slope materials containing at least small fractions of fines. The first movement is usually in the form of a slide failure. The water is supplied from extreme rainfall or in some cases rapid snowmelt. The points of failure are generally located to concave

slope segments, which concentrate surface runoff and interflow.

On both the local and the regional scale, the spatial distribution of debris flows may show an irregular pattern, which can be attributed to local terrain conditions, particularly slope gradient and geometry, or to frequency and localization of major rainstorms. It seems likely that the observed distribution of debris flows presently is the result of repeated events in postglacial time, rather than an intense limited period of slope instability after the deglaciation of the area. Several areas have been affected by debris flow activity on numerous occasions during the last 2000-3000 years, according to lichenometric datings. Other areas may not have been subject to similar extreme rainfall events, or have generally too low slope angles or a lack of loose material to allow debris flows to be initiated.

Individual debris flows of the hillslope type transported between 10 to 10 000 m³ of slope debris in the Nissunvagge area in one day in 1979. The wide range in magnitude with a higher frequency of comparatively small flows is probably typical of large events of meteorological background (cf. Larsson 1982, Innes 1985b). The total amount of debris transfer in this single event of widespread slope instability was of the magnitude of 134 000 m³, accomplished by about 300 debris slides and flows.

As a morphological result of the local downslope transport of material by debris flows, medium-scale landforms of variable persistence are formed, notably debris flow levees, lobes and cones. Mass transfer by debris flows may lengthwise exceed that of rockfall considerably. Scree slopes affected by debris flows therefore develop concave basal segments. The range of debris flows onto gentle footslopes makes them a potential hazard in high relief areas.

5. Slush avalanches

5.1 Introduction

Slush avalanches (slush flows) are rapid mass movements of water-saturated snow (slush), known mainly from arctic areas (Nobles 1961, Washburn & Goldthwait 1958, Mellor 1978). Descriptions of slush avalanches from these regions have dealt with their role in stream breakup and glacier snow ablation (Washburn & Goldthwait 1958, Nobles 1966), their geomorphic impact (Rapp 1960b, Jahn 1967, Washburn 1979) and influence on vegetation (Rapp 1971). Some of these authors also mention the obvious hazard associated with slush avalanches (Rapp 1966, Jahn 1967). In the Swedish mountain area, near accidents have occurred recently (Nyberg 1982). Only limited research has been done on slush avalanches, compared with the comprehensive literature on snow avalanches or debris flows.

Like debris flows, slush avalanches may cause damage to construction works, such as roads (Larsson 1974). Their erosional power can be very great and huge amounts of blocks and earth debris are often incorporated in the mixture of snow, ice and water of the avalanches. Therefore, they have the potential of being a significant geomorphic process in arctic areas, as indicated by the study of Rapp (1960b) from the Kärkevagge valley in northern Sweden. The work by Rapp initiated further studies of slush avalanches in limited areas of Swedish Lapland (Larsson 1974, Nyberg 1980, 1985). A few references to slush avalanches in other parts of northern Scandinavia have also been made by other workers (Söderman 1980, Hestnes & Lied 1980).

The restricted information available for this avalanche type and its related landforms up to the present can be gauged from the omission of it on the recently issued geomorphological maps of the Swedish mountain area. Thus the knowledge concerning its distribution regionally and relationship to terrain characteristics has been even less than for debris flows. Apart from the isolated case studies mentioned above, no information has existed on which to base an evaluation of triggering release conditions, erosional and landforming capacity and temporal frequency of slush avalanches in northern Scandinavia. Hence, from both a geomorphological standpoint and with regard to the potential hazard, further research on slush avalanches is highly motivated.

5.2 General characteristics of slush avalanches

5.2.1 Definition and occurrence

Slushflows were originally described by Washburn and Goldthwait (1958) as 'mudflowlike flowage of water-saturated snow', induced by 'intense thawing, producing more meltwater than can drain through the snow'. They are 'the characteristic method of stream break-up' in NE Greenland, but also occur on glaciers (Op.cit. p.1657-8). A later definition by Washburn (1979, p.193) is 'the predominantly linear flow of water-saturated snow'. He stresses the transitional character of slushflow between wet-snow avalanching and fluvial action, also mentioning the similarities to

debris flows (Op.cit. p.194). The terms slush flow and slush avalanche have been used as synonyms. Also the term 'wet snow avalanche' sometimes appears in the literature presumably with the same meaning, although it could better be reserved for avalanches with lower content of liquid water, lacking the water-saturated snow referred to as slush. There may of course exist gradational forms which complicate this distinction. Furthermore, the possibility exists that some events referred to as 'spring floods' and designating the catastrophic bursting of snow jams at outflow channels of arctic lakes (e.g. Woo 1980), could equally well be termed slushflows.

A distinction is suggested restricting the usage of the term slush avalanche to major events, while slushflow could be applied also to small-scale stream breakups, where the snow is converted to slush.

5.2.2 Snow cover instability and types of avalanches

The literature on snow avalanches is voluminous due to their role as natural hazards in many mountain areas. For detailed information, the reader is referred to the literature, e.g. Perla & Martinelli (1975), Mellor (1978), Perla (1980), Ramsli (1981), Schaerer (1981). A publication summing up knowledge of avalanches in the Swedish mountain area was recently issued by the National Environment Protection Board (SNV 1983). The following general information on snow characteristics is based on the mentioned authors and Gray & Male (1981).

A. Metamorphism of snow

Newly fallen snow undergoes gradual change in response to prevailing weather conditions and as a result of compaction. This change, affecting snow crystal form, is referred to as snow metamorphism and plays a fundamental role in influencing the risk of snow failures. Winter snowpacks are mainly affected by two types of metamorphism, TG- and ET-metamorphism, depending on whether the changes are related to a temperature gradient within the snowpack or not. The former type, temperature-gradient (TG) metamorphism, leads to a growth of individual snow crystals when vapor passes from warm to cold parts of the snowpack through the process of sublimation (Langham 1981 p.277). It is also referred to as 'constructive metamorphism'. Large striated prism- or cup-shaped grains known as depth hoar are the ultimate stage in TG-metamorphism. Since vapor in this process is mainly deposited on individual grains and not on contact surfaces between grains, this type of snow texture tends to be poorly bonded and of low strength. Snow failures are accordingly often initiated in layers of well developed TG-grains.

The second type of metamorphism affecting the winter snow cover is equitemperature (ET) metamorphism, also known as 'destructive metamorphism'. In general terms, it can be described as a 'tendency of the snow to 'simplify' its form and knit into a tighter, stronger texture' (Perla & Martinelli 1975, p.43). Again, this process probably works through transfer of water vapor within the snowpack as a result mainly of sublimation, see further e.g. Perla & Martinelli (1975), Langham (1981). Uniform grain-size snow with strong bonds between grains is the end result of this type of snow metamorphism. It dominates snow development especially shortly after deposition.

In spring, and also during maritime winter conditions, a third type of snow metamorphism becomes active. This meltfreezing (MF) metamorphism, working through differential melting and refreezing of grains of different size, results in the development of coarse-grained snow with large aggregates of individual grains. The same outcome has been observed for snow immersed in water, where particles smaller than about 0.2-0.3 mm are eliminated by metamorphism in 3-6 days (Wakahama 1968). The type of snow formed by MF-metamorphism dominates the snow cover in spring and is known as corn snow. When refrozen, it may have an enormous strength, but as melting occurs, almost all strength disappears and the grains are connected only by surface tension of the present liquid water and may be separated completely by pore water pressures. In this state, the snow is unable to support the weight of a skier, and it is frequently referred to as 'rotten' snow, for reasons easily understood.

B. Wind transport and packing

With respect to avalanches, wind effects on snow cover thickness of particular significance are the building-up of cornices to the lee of crests of steep slopes and the infilling of stream gullies by thick snowpacks, creating snow jams that may obstruct melt water runoff in spring.

Strong winds also affect the structure of the snowpack by increasing density in windblown deposited snow compared to snow deposited by gravity fall alone.

C. Causes of instability of the snow cover on slopes

Depending on the history of a snowpack, that is, its mode of deposition, original snow type stratification, secondary wind influence, and type and degree of metamorphism, the snow has a certain strength, related to cohesion and friction between particles. The strength may vary markedly between different layers, usually linked to grain size, with high strength in fine-grained old snow, low in coarse and new snow.

A snow cover on a slope is subject to gravitational stresses, acting downslope. The strength properties of the snow counteract these stresses. The situation is analogous to the one described for soil-covered slopes regarding stability conditions, expressed by the value of the safety factor, F (section 4.2.3). When imposed stresses exceed existing strength of the snow cover, failure occurs, which may develop into an avalanche (analogous to a slide/flow initiation in earth materials).

Being near its melting point, snow, like other solids, exhibit viscoelastic mechanical behaviour, except for fast, strong stresses, which cause brittle fracture (Perla & Martinelli 1975, p.52). In hard, high density snow, brittle fracture may separate large slabs of cohesive snow along cracks in the snow cover. These slabs may start gliding downslope, usually on a glide plane within the snowpack, disintegrating into blocks of hard snow as movement accelerates. This is the most dangerous type of snow avalanche, the slab avalanche. In loose, cohesionless snow, failure occurs when the critical angle of repose for the snow particles is exceeded. Movement starts point-wise and the mass incorporates progressively more snow downslope. This second main type of avalanche is the loose-snow avalanche.

Very little is known about the release conditions of slush avalanches, but it seems that they sometimes originate as slab avalanches, and the snow is subsequently converted to slush in a stream (Rapp 1960b). At other times they start as a loose-snow avalanche, where a local water table develops in the snow cover (Nobles 1961), and cohesion between snow grains is lost due to excessive pore water pressure. The present study has been aimed at studying the release conditions for this avalanche type in the Abisko mountains. The two forms of release for slush avalanches will be referred to as 'slab' and 'water saturation' release.

As with debris flows, the initiating factors for snow failures may be divided into external and internal factors (cf. section 4.2.3). External factors include increased load due to snowfall, a falling cornice or a skier, adding to the stresses acting on the snowpack; internal factors cause snowpack strength reduction and include snow metamorphism and melting.

5.2.3 Terrain localization

Whereas snow avalanches occur both on steep slopes in general (unconfined) and in steep gullies on mountain walls, slush avalanches in Lapland, have mainly been reported from stream courses. Descriptions of slush avalanche sites along stream courses are few (Rapp 1960 p.138-144, Jahn 1967 p.221-224, Larsson 1974 p.7-17, 22-25). They usually emphasize the importance of a collecting basin as a source area of large amounts of melt water, and some type of barrier at the outlet of the basin, e.g. a snow jam or a frozen water-fall, whereby damming of melt water and saturation of the snow cover can occur (cf. also Cottman 1966). Slope gradients of the starting zones may be quite low compared with snow avalanche sites. Luckman (1977) cites values of 10-15°.

A second localization described in the literature is on glacier surfaces, where part of the melt season runoff from the soaked zone (slush zone) can take the form of slushflows (Nobles 1961, 1966). The slope gradients necessary for the release of slush avalanches on glaciers can be very low (less than 10°) and motion of slush has been observed on gradients of less than 2° (Op.cit.). Since there are no direct geomorphic effects of slush avalanches on glaciers, they are not dealt with in the present study.

Other terrain localizations have been reported, also from areas outside the Arctic, such as on talus slopes and in shallow depressions along sloping valley bottoms, although the descriptions given do not definitely establish identical process characteristics for these cases and the Arctic ones (Caine 1969, Bones 1973, Starkel 1976, Hestnes & Lied 1980, Fort 1981).

Further data on site characteristics of slush avalanches appear in section 5.4.

5.3 The release conditions of slush avalanches

5.3.1 Sources of liquid water in the snowpack

Liquid water in a snowpack arises from two sources: snow melting and rainfall. Studies of snow melting from a hydrological viewpoint are numerous, e.g. Forsman (1963). Melting is caused by energy inputs at the snow surface, or less important, from the

underlying ground (Q_g). Energy sources at the surface, ultimately derived from the sun, are a/ short and longwave radiation (Q_{sn}, Q_{ln}), b/ convection of warm air (Q_n) and c/ latent heat as phase changes at the snow surface (Q_e). Rainfall contributes marginal energy (Q_p) to snow melting, but is primarily important as a second, direct source of liquid water to the snowpack. The equation describing energy inputs for snow melting is thus (Male & Gray 1981, p.362):

$$Q_m = Q_{sn} + Q_{ln} + Q_n + Q_e + Q_g + Q_p - dU/dt$$

where Q_m =energy flux available for melt, dU/dt =rate of change of internal energy of the snow cover (e.g. rewarming after nocturnal freeze before melting can start)

Liquid water supplied by melting from radiation is usually most abundant by midafternoon on a diurnal basis. This relationship is caused not only by the delayed effect of maximum insolation but also by the fact that the snow cover becomes increasingly potent in absorbing radiation as it gets wetter (lower albedo). Spatial variations in liquid water content are caused by local increase in melt rate resulting near rock walls, protruding blocks etc. due to longwave radiation (heat transfer) from these objects, and by flow concentration to hollows or along impermeable strata. Both temporal and spatial variations in liquid water content in a snow cover may be of significance in the timing and localization of slush avalanches.

5.3.2 Retention and percolation of water in the snowpack

Liquid water in small amounts (initial melt stage) occurs adsorbed to snow grains. With continued melting, the water starts flowing through pores or along drainage pathways, initially by capillary forces, later (above about 1% water by volume (Langham 1981 p.325)) governed by gravity. A water content high enough to create a continuous liquid matrix surrounding the grains of the snowpack is known as the funicular regime. This condition covers the range from about 7-12% up to full saturation of available pore volume for old, coarse-grained snow (Denoth 1982). At lower saturation levels, a discontinuous water film is found around grains, with gaseous paths in between, a condition referred to as the pendular regime. Capillary pressures and bonding strengths between grains are considerably higher in the pendular compared to the funicular regime (Male 1980, p.372). Most springtime snowpacks have free water contents in the pendular regime (Op.cit.).

With further increase in liquid water content to more than about 25%, the snow converts to a slushlike consistency (Perla 1980, p.401). Meltwater runoff along the ground begins once the whole snowpack has been warmed up to 0°C (the snow is isothermal) and the infiltration capacity of underlying soil is exceeded, if not frozen. Before this happens, much of the surface-produced meltwater refreezes on its way downwards through the snow and contributes, by release of latent heat, to the warming-up of the lower cold layers of the snow. In this process planar ice crusts are also created, which may act as impermeable strata to subsequent meltwater percolation. Hence, water becomes concentrated in certain horizons, until weakening of the layers opens up vertical drainage channels for downward percolation. This type of water retention along hard layers may give rise to planar weaknesses in the snowpack, which can act as slide planes for avalanches

(Perla & Martinelli 1975, p.51). Sometimes layers of slush are formed above impermeable crusts in a snowpack.

5.3.3 The influence of liquid water on snowpack stability

A review of research on water production and movement in snow is given by Male (1980).

Liquid water in such amounts that are not held adsorbed to snow grains but starts moving by gravity forces in a snowpack, often referred to as free water, has a profound effect on snowpack stability. Increased liquid water content causes higher compressibility of snow (Colbeck & Parssinen 1978). The pore water pressures eventually loosen bonds between grains, thus drastically reducing cohesive and frictional strength.

The negative influence on snow strength by increasing amounts of liquid water was demonstrated by Ambach & Howorka (1966), who found a clear relationship between incidence of wet snow avalanches and content of free water in the snow cover in the Swiss mountains. They noted significant synchronicity in the variations of daily mean free water content and mean air temperature. During the study period in April-May 1962, avalanches occurred as free water content exceeded about 7 vol%. No values higher than 10% were recorded. Moskalev (1966) states values of 10-15% as critical for wet snow avalanche initiation. He analyzed the stability conditions of wet snow on slopes and stressed the importance of pore water pressures and water seepage in basal aquifers in the snowpack. Increasing thickness of a basal aquifer (slush layer) will lead to potentially unstable conditions on successively lower slope angles. Initially, snow melt runoff beneath a snowpack takes place in a saturated slush layer of shallow depth (Colbeck 1974), with little direct effect on overlying snow strata. However, with increased melting and discharge of flowing water, the thickness of the slush layer gets larger (Op. cit.). If the entire snowpack becomes soaked with water and converts to slush, the general loss of strength between snow grains, due to pore water pressures and melting of bonds between grains, may lead to release of slushflow. Free water content as high as 35-50 % has been suggested for slushflows (Nobles 1966, Mellor 1978). This situation probably only occurs under conditions of impeded drainage of melt water from the snowpack, e.g. in topographic situations where snow or ice jams prevent free drainage downslope in connection with rapid thaws. Several factors determine the potential of snowmelt in producing high inputs of water liable to result in damming (cf. Ward 1980 p.28). A rise in temperature will have variable influence on snowmelt production depending on snowpack characteristics, e.g. whether the snow initially is cold or isothermal (at melting point). Properties such as porosity, permeability and presence of ice layers govern meltwater movement through the snowpack and thus determine whether runoff will be direct or with a time lag. Furthermore, substantial damming of meltwater will result only when the ground surface is frozen or has been saturated.

From these basic considerations, we will turn to some experiments and observations on actual snow conditions at slush avalanche sites.

5.3.4 Field experiments on the strength of wet snow

Detailed consideration of the physical mechanisms and controls responsible for the release of slush avalanches has been outside the scope of the present study. However, to obtain some quantitative data on the relationship between content of free water in the snowpack and its strength, field measurements were carried out in the Abisko area. The solution method and Ram penetrometer used have been described in chapter 3. The measurements were made at Mt. Njulla and in Kärkevagge. The procedure included repeated measurements of snow strength and water content after successive sprayings with water on a series of adjacent squares (1m^2), each receiving different amounts of water. The snowpacks were isothermal at the beginning of the experiment and made up of corn snow with a basal layer of TG-metamorphosed grains. The results of the experiment are shown in Table 5.1 and Fig. 5.1.

A. Discussion

Variations in water percolation could produce the spread in measured values. The rate of percolation was measured by injection of a dye solution and found to be about 4 cm/min in June 82 and slightly higher in May 83. The values fall within the lower range of generally observed rates of 2-60 cm/min (Male & Gray 1981, p.398). The percolation rate presumably increased gradually due to formation of pathways for water in the snowpack. These are also thought to explain why successively higher water contents were not obtained in the tests, but the content seemed to fluctuate between 12-21% in the first run and stabilize around 11% in the second (Table 5.1). A laboratory sample of snow immersed in water that may represent complete saturation of available pore space with water, gave a value of 58% for free water content. In the absence of any factor causing impeded drainage, it seems that values of free water content intermediary to those achieved in the experiment and fully saturated values are unlikely to occur in a snowpack. Forsman (1963, p.40) states that about 8% liquid water by weight represents a limiting value that the snowpack will hold, before basal runoff begins. In an experimental series of relevance in this context, Gerdell (quoted in U.S. Army Corps of Engineers 1956) obtained peak values of about 10% liquid water content by weight, with a maximum of 16% after application of water to snowpacks of densities $350\text{-}460\text{ kg/m}^3$.

The influence of increasing free water content on Ram strength of the snowpack for Kärkevagge (by the hut in the valley bottom) on May 11, 1983 is shown in Fig. 5.1. The snow had the frequently observed contrast in strength between the middle and low part of the profile, due to metamorphism during the winter season. With successive additions of water to the snowpack, the uppermost weak layer grew progressively downwards and lost further strength. The strong central part was eventually eliminated and the penetrometer sank all through the snowpack almost by its own weight. The strong basal layer in the final profile was probably a consequence of compaction of icy wet snow by the instrument and not formed as a result of the increased water percolation.

Some further measurements were carried out to study naturally occurring contents of free water in spring time snowpacks (Table 5.1). The values at Mt. Njulla refer to a warm melt situation of May 1983 producing surficial slush along the Ridonjira brook

Table 5. 1. Measured content of free water in snow

<u>Field experiments</u>			<u>Laboratory sample</u>	free water vol%
	mm water added	free water vol%	water-drenched snow (slush)	58.6
<u>Field samples</u>				
Kärkevagge 20 June			Njulla, along brook	
20 June 1982	0	5.0	upslope of waterfall (16. 5. 83)	
infiltration 7 min	2	1.8		
	4	15.0		
	6	21.4	surficial snow above slush layer	10.0
	8	15.5	slush, drained of water	17.6
	10	12.3	deep snowpack 20 cm depth	16.4
	12	21.3	" " 100 "	16.5
11 May 1983	0	8.1	Abiskojokk, along the shore	
infiltration 4 min	2.5	8.5	(13. 5. 83)	
	5	10.9	water-soaked snow	29.1-31.2
	7.5	11.3	Kärkerieppe cirque, near	
	10	15.0	threshold (10. 5. 83)	
	12.5	11.2	deep snowpack 40 cm depth	2.5
	25	11.5	" 100 "	2.2

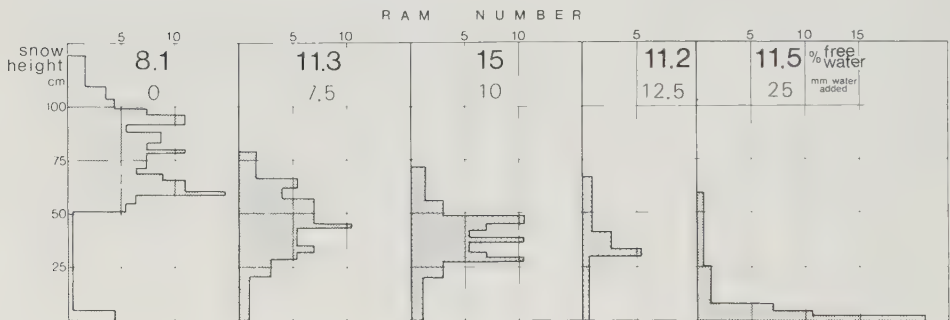


Fig. 5.1. Snow strength, measured as penetration of a Ram sond, in relation to liquid water content of the snow, measured by the solution method. Increasing amounts of water were added to the snowpack artificially, and Ram strength measured after the water had infiltrated for 4 minutes. Samples for determination of liquid water content were taken at a depth of 40 cm below the surface. The fairly poor correlation between the added amount of water and the measured liquid water content was presumably due to the development of efficient drainage paths through the snowpack. Kärkevagge, May 11, 1983. Cf. Table 5.1.

upslope of the steep waterfall. The low water contents found in the Kärkerieppe cirque may be due to low air temperatures, giving minimal melting on the occasion of the measurement.

B. Concluding remarks

The content of free water in a snowpack is no doubt related to the potential for slush avalanching, although absolute critical values can hardly be given. A very complex interaction of factors is likely to be involved in the release of slush avalanches. The interrelationships are indicated by the scheme of Fig. 5.2, showing some of the presumably most important factors. The primary concern with regard to water-saturation release of slush

avalanches is to detect incipient damming of meltwater in a snowpack, which demands excavation of snow pits at potential sites to be made, unless more sophisticated methods can be applied. Among these, tensiometers, some type of water level gauge and microwave sensors seem plausible (cf. Tobarias et al. 1978, Wankiewicz & de Vries 1978).

5.3.5 Snow conditions at slush avalanche sites in the Abisko area

During the years 1979-1983, snow conditions were studied at two sites for slush avalanches, one with a westerly location in partly maritime climate (Kärkevagge) and the other in a more easterly locality of partly continental climatic character (Kaisepakte). These sites had large slush avalanches in 1980 and 1979, respectively. The purpose of the studies was to obtain information on snow depth and type of stratification prior to and during melting at starting zones of slush avalanches, especially if damming of melt water occurred, and whether basal ice crusts developed on the stream beds. Examples of snowpack conditions at the two sites are given below. A more exhaustive presentation of the results for the Kärkevagge locality is included in Appendix 2. In addition, the snowpacks in starting zones of four other sites, where slush avalanches occurred during the study period, were investigated.

The situation of the studied sites and other sites in the Abisko area is shown in Figs. 5.3 and 5.4.

A. Kärkerieppe, Kärkevagge.

The site is situated at an altitude of about 1000 m along the stream from the NE-facing hanging cirque Kärkerieppe. A small remnant of the cirque glacier and large snow fields supply the stream with melt water from the end of May until September. The stream channel, which is incised down to bedrock along parts of its course, is regularly filled with deep snowpacks at the beginning of the melting season.

The study site comprised two transverse profiles upslope of the lower rim of the cirque, one (A) across a gently sloping (about 10°) 10 m wide tributary stream channel 50 m upslope of the main stream gully from the cirque, and the other (B) about 50 m to the north of profile A, on bedrock outcrops near the main channel (Cf. Fig. 5.5). The channel bed of profile A consists of till with numerous stones and a few boulders.

Snow depth

Values for snow profile A1 (Fig. 5.6) are given in Table 5.2. Due to weather conditions and also for practical reasons, the measurements refer to different dates in the various years. Generally, they indicate approximate maximum amounts of snow, prior to spring melting (except for 1980 when bad weather conditions precluded measurement in May). During the research period, 1981 and 1982 had the largest snow depths, while 1980, from the only measurement made, appears to have been a year of comparatively

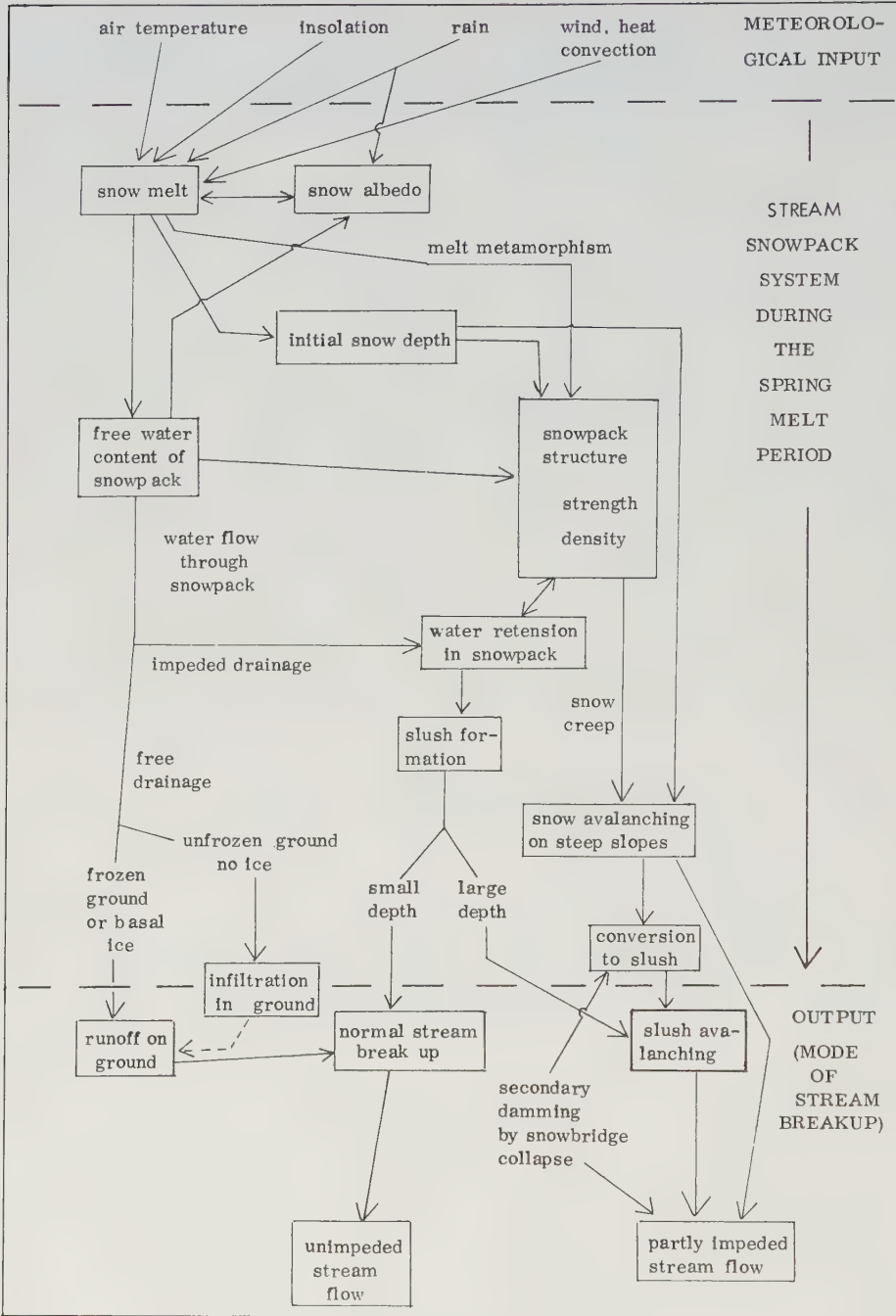


Fig. 5.2. Outline of factor interactions involved in the release of slush avalanches in stream courses. The diagram is structured into three levels: the external factors of meteorological origin, the internal factors relating to the stream bed snowpack system, and the resulting mode of disintegration of the snowpack during the snowmelt period in spring. Slush avalanching is shown as an alternative way of stream breakup, conditioned by factors such as snowpack drainage/water retention, snowbridge collapses or snow avalanching in steep sections. Avalanches influence the flow regime of a stream, giving irregular discharge with time lag effects.

small depths. Similar variations occurred for the nearest weather station at Katterjåkk (Fig. 5.7).

The general snow accumulation pattern for the cirque threshold was estimated from a survey in May-July 1982, using fixed stakes and intermediate soundings along several transverse and longitudinal profiles. The largest thicknesses were found along the northern tributary (over 2 m depth). Lower values were located to the interfluvial areas (slightly under 1 m). Snow depths of the main stream gully were not investigated for security reasons, but are presumably considerably larger than on surrounding slopes and the cirque bottom. The differences are due to wind redistribution of the snow cover.

Snow stratification

Examples of the snow pit studies at the Kärkerieppe locality are shown in Fig. 5.8. The profiles from 1981 display a transition from early melt conditions (b), with fairly large variations in Ram strength in different strata and a well developed weak basal layer of advanced TG-metamorphism crystals (depth hoar), to late melt conditions (c), with generally low Ram strength values and reduced contrasts between strata, which predominantly consist of coarse, MF-metamorphism grains. Both profiles were isothermal. In profile a, a late winter situation from April 30, 1982, influenced by some melting, is shown. The snow cover displays a weak vertical temperature gradient. Strength is higher than for the melt season profiles. A weak basal layer is established also this year. Basal ice crusts were not present in these profiles, but have been observed in other cases (see Appendix 2). Each year thick ice crusts covered the lower part of the stream course, upslope of the alluvial fan apex, probably partly formed as the stream froze up in the early winter. This extrusive ice could be referred to as a small form of stream naled, although it melts away early in the summer (cf. Åkerman 1980, 1982). Its position being at the foot of the stream gully, it has no significance with respect to the release of major slush avalanches.

Snow melting

The broad progression of snow melting at the site was followed via the repeated snow depth measurements annually. The timing of the snow melt during the years of the study period fell between the end of May to beginning of July, except in 1982, when it lasted to the middle of July.

Diurnal snow cover ablation rates from 1.3-9.8 cm (about 5-39 mm water equiv.) were measured in 1982. Data recorded by the THG of a meteorological screen by the hut in Kärkevagge valley bottom indicate that effective melting only began towards the end of June this year at the site. Practically no runoff occurred in the stream from the cirque before this time, as shown by readings of a stream gauge in the valley bottom.

Formation of snow slush caused by damming of meltwater was observed in 1979, when 70 cm of water and slush was found beneath 95 cm of snow at profile A. The wetness of the snow on other occasions, judged by the hand test, was generally 'moist' to

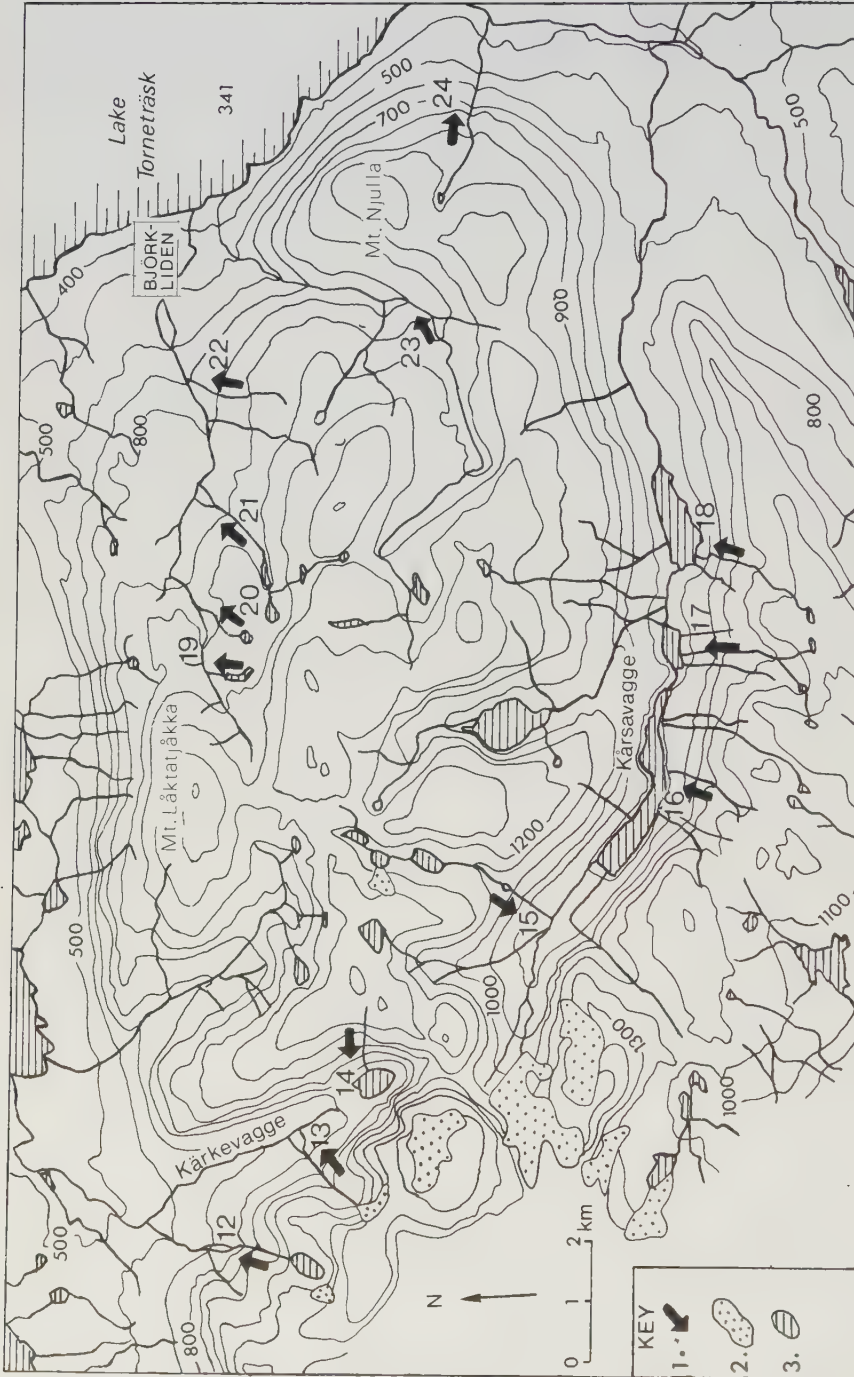


Fig. 5.3. Distribution of slush avalanche sites along stream courses in the western Abisko high mountains. Key: 1. slush avalanche site 2. glacier 3. lake. Note the predominating north to northeast orientation of sites.

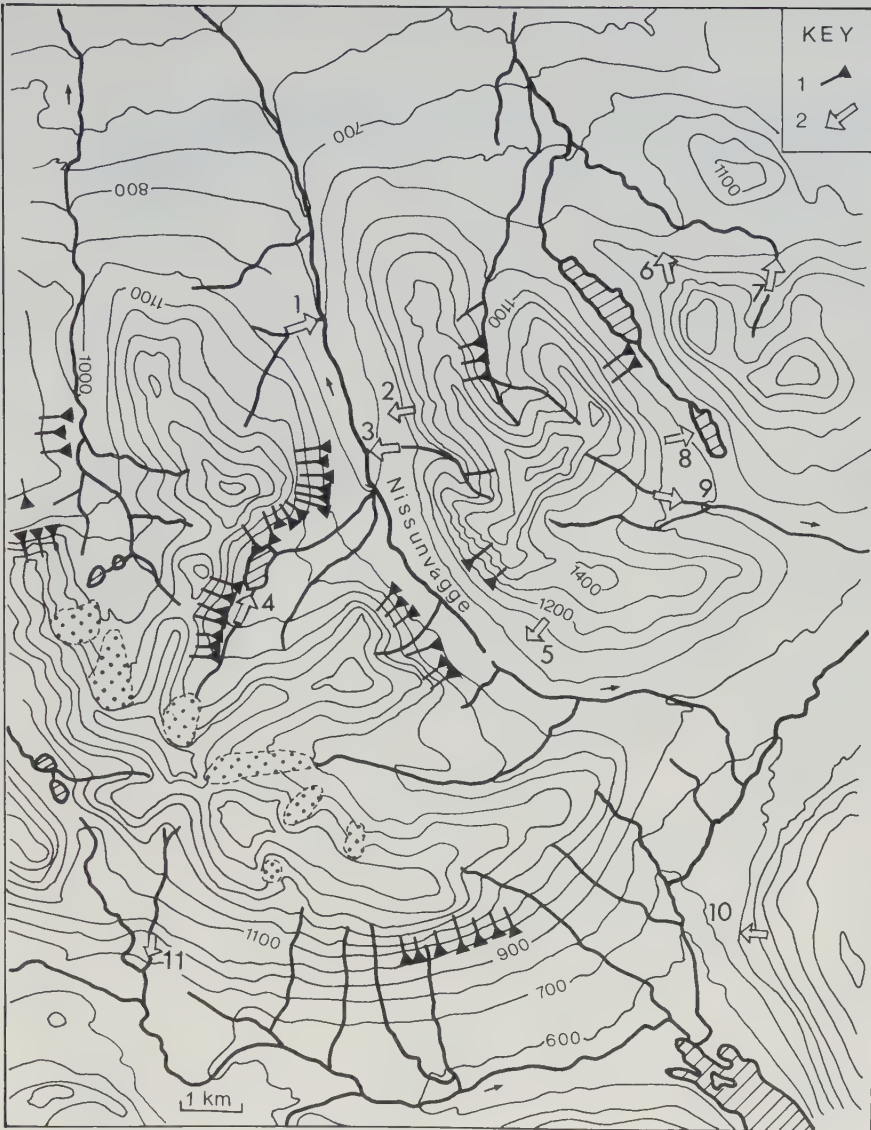


Fig. 5.4. Distribution of slush avalanche sites along stream courses and snow avalanche sites of geomorphological importance in the southern Abisko high mountains. Key: 1. avalanche boulder tongue due to snow avalanches 2. slush avalanche path. Glaciers shown stippled. Heights in m. a.s.l. Note the different localization of snow and slush avalanches, the former occurring in clusters on steep rockwalls, the latter isolated along stream courses (the smallest streams not shown).

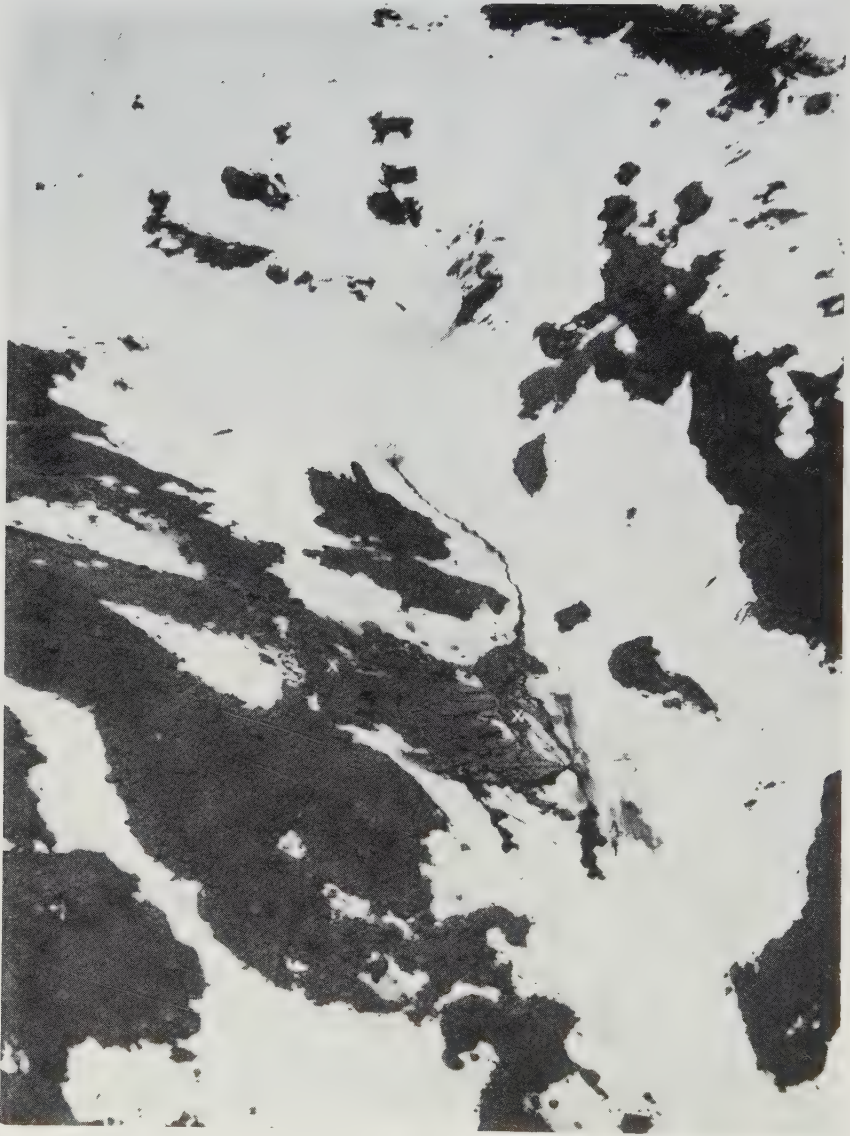


Fig. 5.5. Lower part of the Kärkerieppe cirque, with present glacier and cirque backwall in the background and terminal moraines and bedrock threshold in the foreground. Two meltwater streams converge towards the rock threshold and the snow-filled stream gully, where slush avalanches occur. The sites for snowpack studies marked with x. Note the debris fans upon the snow cover below the rock threshold, in the extension of the two melt water streams. They are indicative of the damming effect of the snow masses filling the gully. Air photo July 12, 1982.

Table 5.2. Snow depths at the Kärkerieppe and Kaisepakte slush avalanche sites (starting zones) during the period 1979-1983.

date	snow depth	date	snow depth, cm
Kärkerieppe 6. 6. 1979	165	Kaisepakte 4. 5. 1981	180
30. 3. 1980	140	8. 6. 1981	> 90
5. 5. 1981	280	29. 4. 1982	>190
6. 6. 1981	140	12. 5. 1983	355
30. 4. 1982	155		
24. 5. 1982	190		
12. 6. 1982	168		
20. 6. 1982	176		
5. 5. 1983	153		
10. 5. 1983	125		
7. 6. 1983	120		

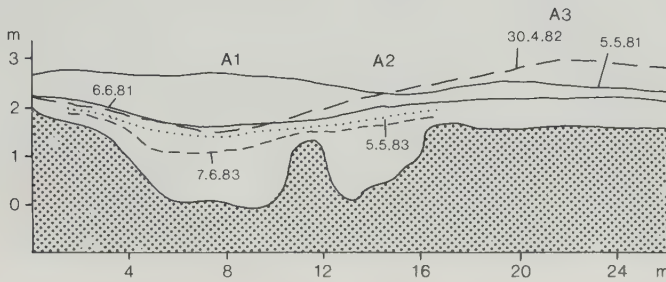


Fig. 5.6. Snow profile A, Kärkerieppe cirque. Measured snow depths during different years show considerable variations, partly due to different snow drift locations as a result of wind transport. The profile is across a small stream channel with a central boulder.

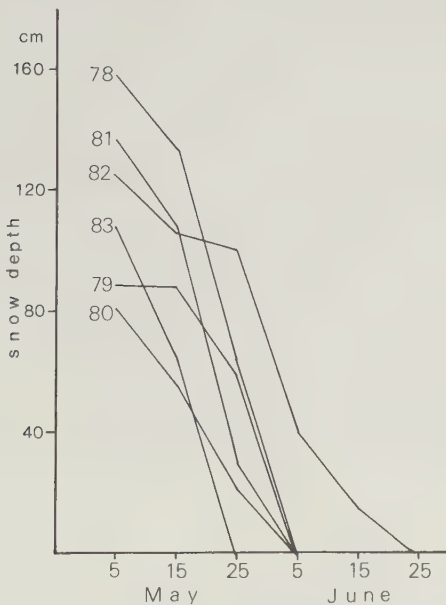


Fig. 5.7. Snow cover ablation during the years 1978-1983 at Katterjåkk.

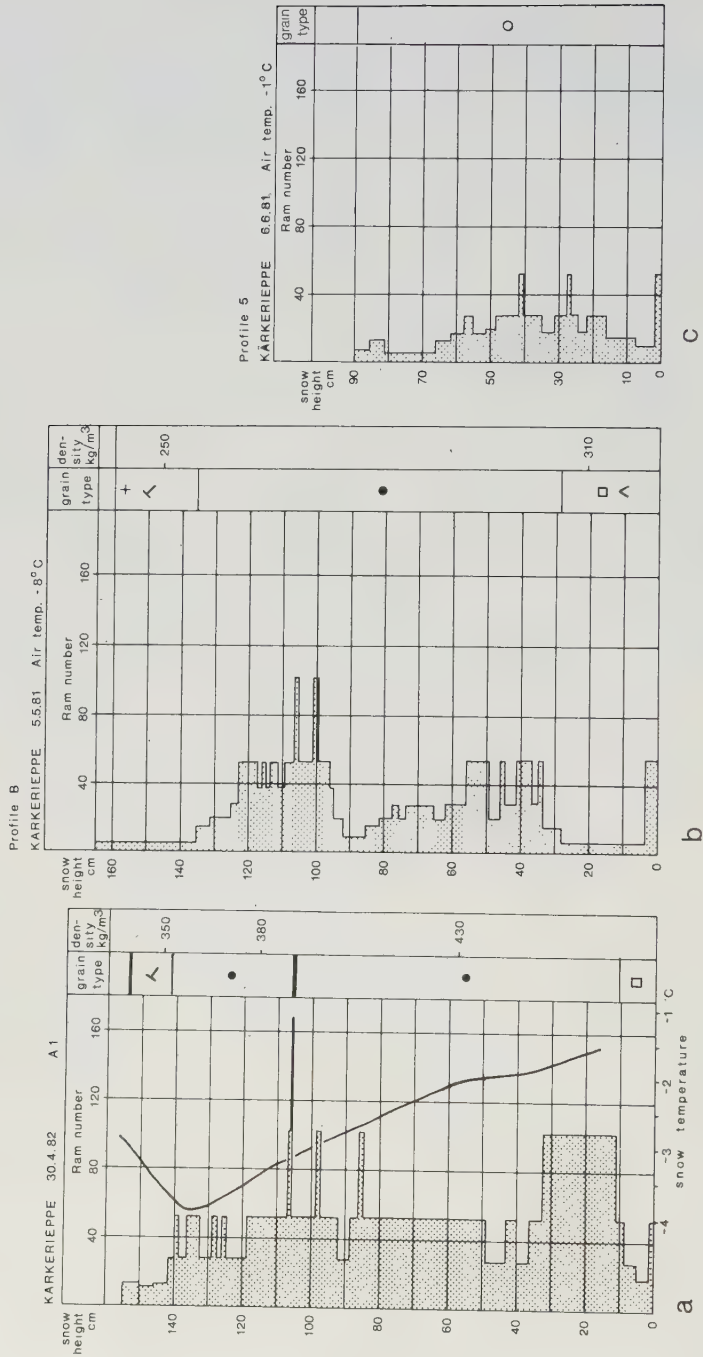


Fig. 5.8. Examples of snow conditions during the late winter or spring period at the slush avalanche site at Kärkerieppe, Kärkevage. Cf. text.

'wet' (2-3 on a scale 1-5, using a standard nomenclature). In some instances basal coarse grained layers were found to be 'very wet' (4). Slush occurred only as very thin layers (1-2 cm) either along sub-horizontal ice crusts or on the ground surface, formed by concentration of slowly flowing water above an impermeable substrate.

On the basis of the available information, it cannot be concluded which type of mechanism for slush avalanches is active at the site, although the observations in 1979 support the possibility of water-saturation release occurring. The generally deep snow-packs and large source area for melt water runoff are factors that may lead to high free water contents of the snow near the cirque rim in years of rapid melting. However, the occurrence of the avalanche in 1980 indicates that the site may be hazardous also in years of fairly thin snow cover.

B. Kaisepakte

The snow study site was a 30 m wide gentle stream course 50 m upslope of the steep canyon Pânjenjâske at about 900 m altitude. At the location of the measurement profile, the stream channel is incised 2-3 m and has a cover of stones and boulders on the bed. The surroundings of the stream course constitute an open, gently sloping blockfield, strongly exposed to wind action. The channel obtains much of its snow cover from wind drifting, with depths far in excess of those found on the blockfield.

Snow depth

Due to the presence of very hard snow or ice layers, it was on several occasions not possible to determine accurate snow depths. The data sampled are given in Table 5.2. Most likely snow depths of June 1981 and April 1982 were substantially larger than indicated. The channel was infilled almost level with surrounding ground, implying average depths of 2-3 m.

Snow stratification

Two situations are exemplified here. Fig. 5.9 shows a penetrometer profile from May 1983 covering 3.5 m snow depth, the highest value encountered in the study period. The melting was fairly early this year, so a general decrease in strength of the snow had begun. However, values were high at this site of strong wind packing of snow compared with those at Kärkerieppe, with less important wind influence. Thin ice layers occurred about 1 m below the surface, with 3-5 cm of the overlying snow transformed to slush. Snow densities were about 500 kg/m^3 . On other occasions, very high values ($600\text{-}700 \text{ kg/m}^3$) were measured.

The other situation exemplifies extremely 'rotten' snow of depth hoar crystals, encountered in the stream channel at 600 m altitude in June 1982.

Snow melting

Generally, snow depths are smaller in this area compared with the more westerly Kärkevagge. Hence, the potential for high melt water production from surrounding slopes is also smaller, which implies reduced chances of water-saturation release of slush

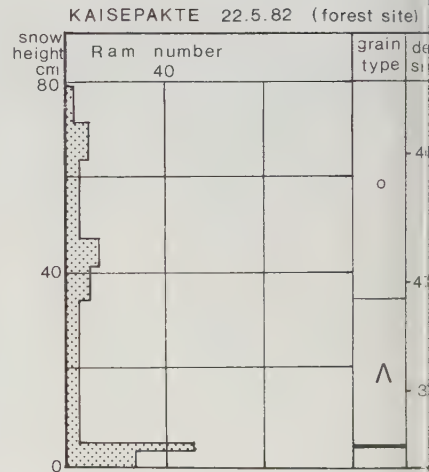
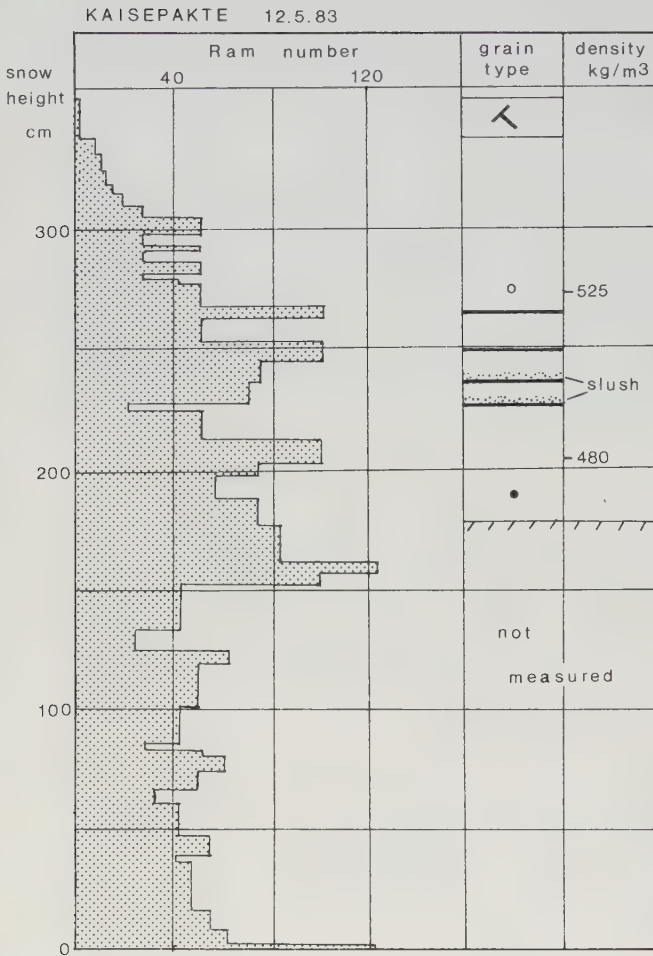


Fig. 5.9. Examples of snow conditions during the spring period at the slush avalanche site at Pånjenjäske, Mt. Kaisepakte. Cf. text.

avalanches. However, as strong winds are important, very deep snowpacks may build up in the stream channel, possibly even in years of generally small snow fall amounts. Thus, snowpacks survive complete melting of the snow on surrounding slopes more frequently here than at the Kärkerleppa site. However, no case of meltwater damming resulting in significant slush formation was noted during the years of observation. The steep canyon head probably favours slab release as initiating factor for slush avalanches at this site, as indicated by observations of the latest avalanche in 1979.

C. Snow conditions at other sites

During a period of steeply rising temperatures in early June 1982, four slush avalanches were observed in the Abisko area. The snow conditions in the starting zones of three of the avalanches were studied about two weeks later. Although the snow cover naturally underwent some changes in the intervening time between avalanche release and inspection, the sampled data afford some indications of the prevailing snow stratification on the occasion of the failures.

Mt. Njulla

The best documented of the cases is that at Mt. Njulla, near Abisko, on 2 June. The avalanche was released around 12 a.m. following a warm spell producing strong melting. It derived most of its snow and slush masses from the snowfilled brook gully of Ridonjira, although probably released as a stream breakup upslope of the waterfall. The range of this avalanche was about 2 km, as it reached the stream Abiskojokk and even crossed this for a short distance. Parts of the birch forest on the margins of the brook were felled (Fig. 5.10). The avalanche carried huge amounts of rock and soil debris, which were deposited along its course and in Abiskojokk. On that occasion, several people narrowly escaped being struck by the avalanche (Nyberg 1982).

The exact location of the starting zone was difficult to determine, when the site was visited on June 15. There were signs of a 'water-saturation' release in the gentle stream course between Mt. Njulla and Mt. Slättatjåkka, where the snow cover was broken and shoved into low lateral ridges. However, most of this slushy snow stopped before it reached the steep waterfall of the Ridonjira brook. Shortly upslope of the waterfall, there was an indication of a 'slab' release, in the form of a 0.5-1 m high scar. A penetrometer profile was measured in the snow cover above the scar (Fig. 5.11).

A 15 cm thick basal ice crust covered the stream bed, with a small air-filled cavity in between. The ice crust partly remained on the parts of the bed that had been cleared out of snow by the avalanche and hence appears to have served as a glide plane for heavy wet snow (cf. Fig. 5.12). The crust may have promoted high pore water pressures in the snow cover by functioning as an impermeable barrier to water percolation. The rise in air temperature preceding the release of the avalanche must have caused a high input of melt water to the snowpack.

On May 16, 1983, the snow melting was earlier than the previous year and the Ram-profile indicated very low strength throughout the snowpack. Only a very thin (2 cm) ice crust was present at the bottom of the snowpack, below which runoff to 6 cm depth occurred (as pure water, running briskly, not as slush). The content of free water in the snow was about 16% (cf. Table 5.1). Upstream the snow cover had broken up in narrow, winding channels, flanked by very low ridges of wet clean snow. The previous year the snow cover was cleared out over a much wider zone along the stream, with higher lateral ridges of dirty snow.

Although no avalanche occurred in 1983, the early melting resulted in a small accumulation of slush in a level section of the stream, some 100 m upslope of the waterfall. The slushy area was



Fig. 5.10 a. Slush avalanche at Ridonjira, Mt. Njulla, near Abisko, June 2, 1982. The avalanche probably started shortly upslope of the water-fall in the background. Its total travel distance was almost 2 km. Several people were present in the immediate danger zone on the occasion, showing the hazard of this little-known avalanche type during the spring season. Photo O. Nordell 2.6.1982.



Fig. 5.10 b. Runout zone of the slush avalanche at Ridonjira, Mt. Njulla on June 2, 1982. Abiskojokk stream in the foreground. Numerous trees were felled by the avalanche along its course. Ridonjira brook gully is seen in the left background, about 1.5 km distant. Photo 10.6.1982.

observed to respond to irregularities in the stream flow, due to minor snow bank collapses upstream, causing temporary dams (cf. Ballantyne & McCann 1980). As a wave of higher water approached, the slush layer (10-20 cm deep) was successively heaved until it suddenly ruptured and released small water flows on top of it. One conceivable mechanism for the initiation of a slush avalanche is where discharge irregularities due to snow bank collapses cause wave fronts large enough to overcome the sluggishness of an accumulated mass of slush, setting it in motion downslope.

In 1984, a medium-sized slush avalanche again occurred at the Njulla site (Fig. 5.13). A comparison of the temperature conditions at Abisko during the critical periods for slush avalanches is shown in Fig. 5.14 covering the years of the study, with observed cases at Mt. Njulla marked. Two of the cases occurred after rapid temperature increase (1981 and 1984). In 1982, the temperature increased more gradually over a longer period. Apart from in 1985, the mean temperatures were generally above the average during the spring period in the years of the study. Variations in snow depths between the different years are also shown. As noted earlier for the Kärkerieppe site, avalanches occur not only in years of thick snow cover.

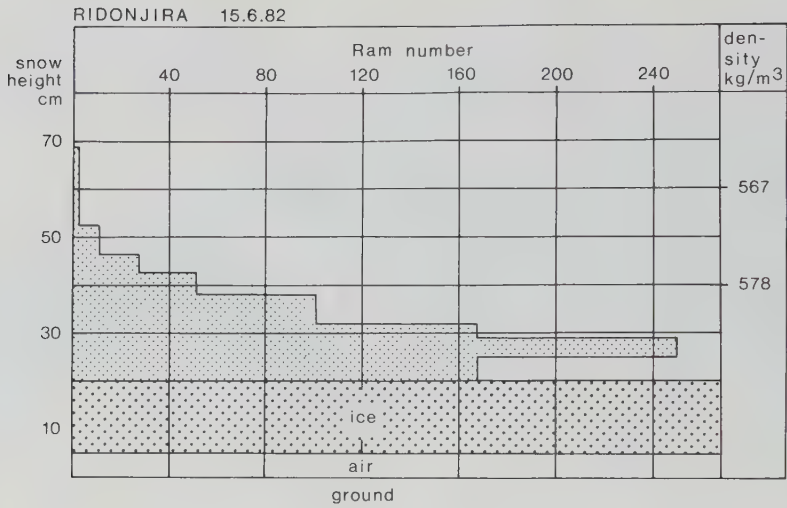


Fig. 5.11. Strength variations measured in the snow cover upslope of the steep fall section of Ridonjira brook, Mt. Njulla, at a point where a one m high vertical scar indicated a slab avalanche release.



Fig. 5.12. Remaining ice and snow layer in the brook Ridonjira after the slush avalanche on June 2, 1982. The striations were formed by gliding debris-rich water-saturated snow. Location near tree line. Photo 10.6.1982.



Fig. 5.13. Slush avalanche at Ridonjira, Mt. Njulla, on May 17, 1984. Note the vertical scars in the snow cover at the waterfall and shortly upslope along the stream, and the remaining snow (or ice) in front of the scars. These features indicate a release as a slab avalanche, probably due to heavy water-saturation of the snow increasing its weight, or to a 'hydro-pressure' effect by subnival stream flow or flow along a basal ice crust. Photo N.Å.Andersson 18.5.1984.

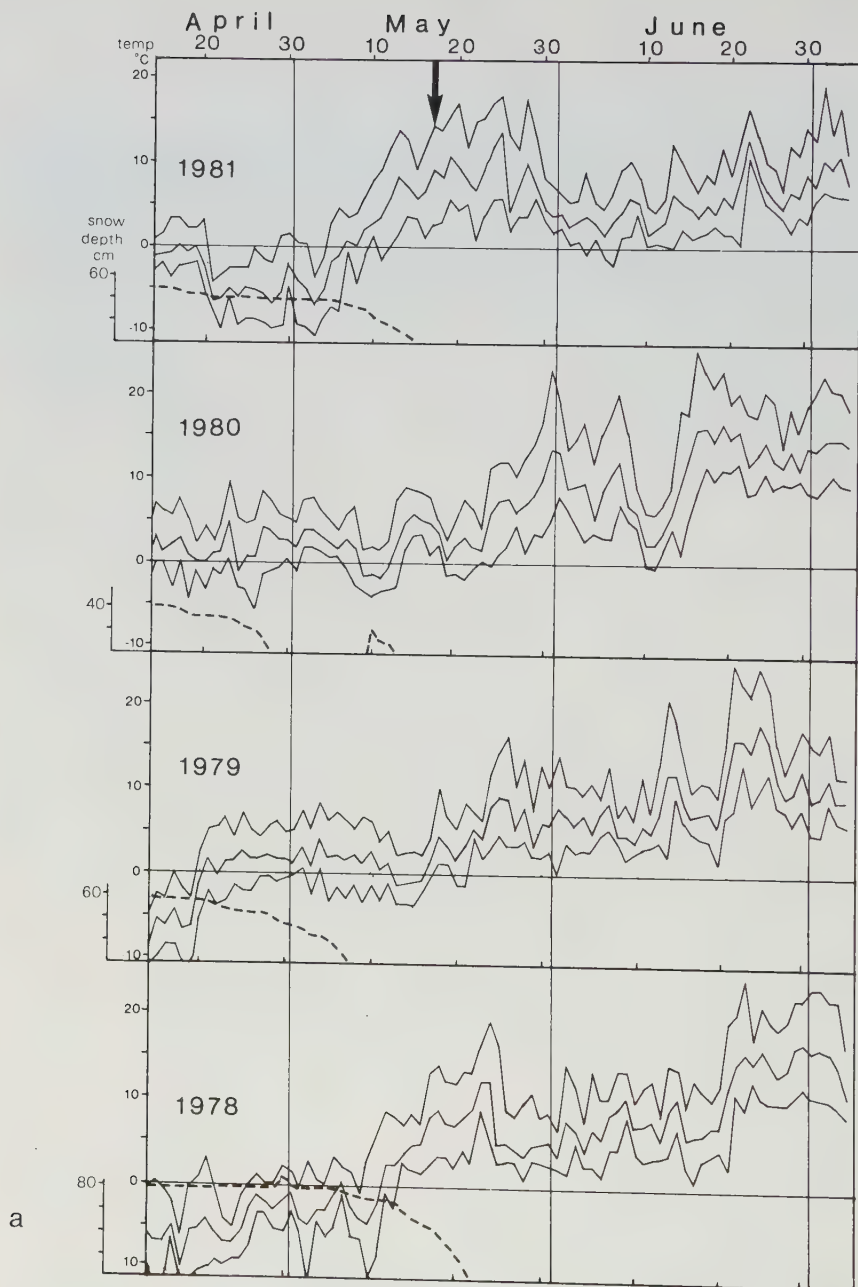
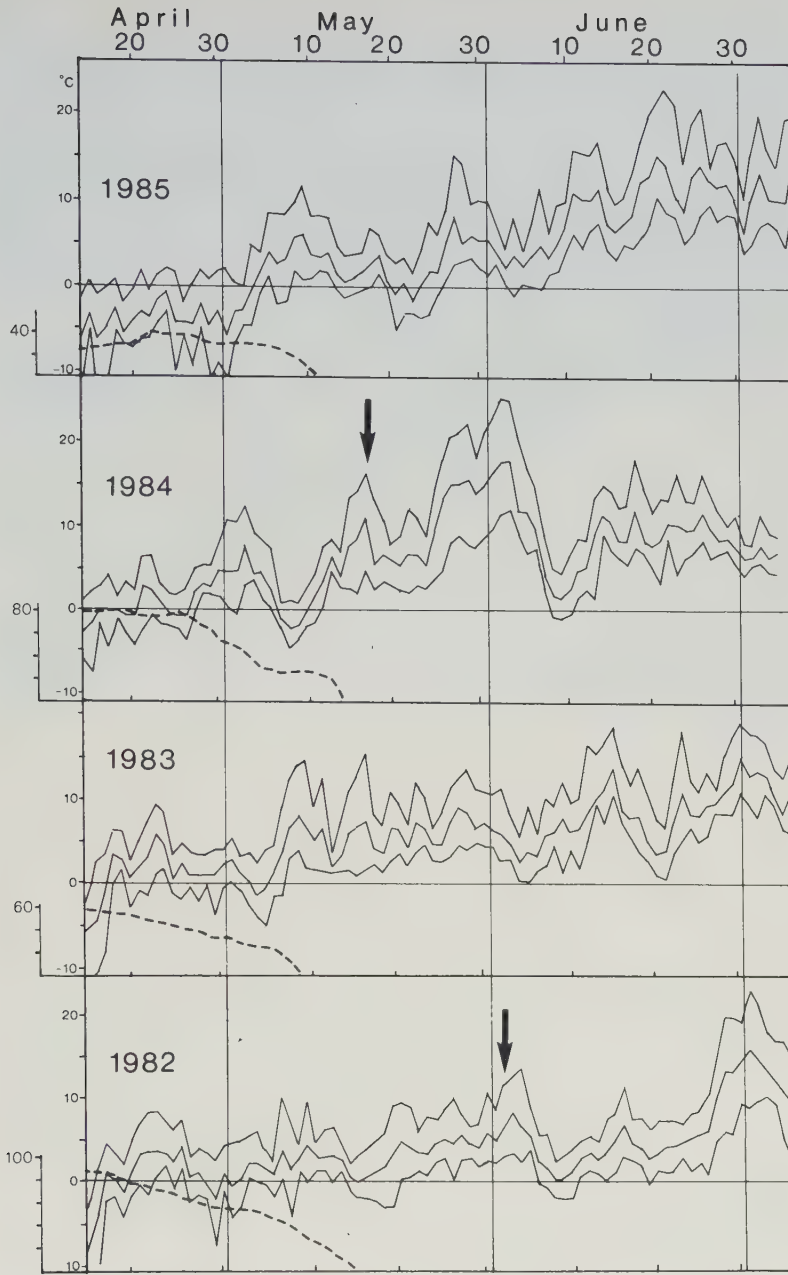


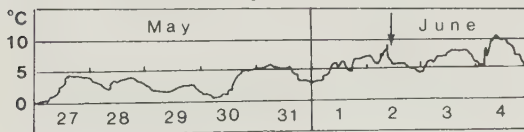
Fig. 5.14. Temperature and snow depth data for Abisko (370 m a.s.l.) during critical periods for slush avalanche release. The cases of slush avalanches at Mt. Njulla, 4 km from Abisko and at about 900 m altitude, marked by arrows. Also shown is a temperature record at the tree line, Mt. Njulla, on June 2, 1982

a-b. Diurnal temperatures and snow depths,
c. mean monthly temperatures during 1978-85.

Data submitted by Nils-Åke Andersson, Abisko and Olle Nordell, Lund.



Tree-line Mt. Njulla



b

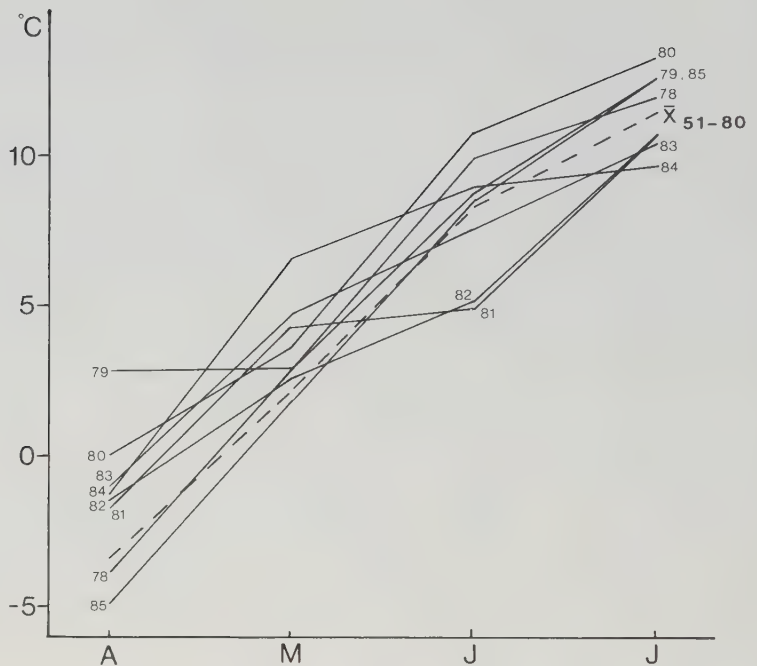


Fig. 5.14 c

Mean monthly temperature at Abisko during the snowmelt period 1978 - 1985, compared with long-term mean (1951-80).

Rakasjokk

The remnants of a slush avalanche was observed on June 14, 1982 in the area west of Björkliden (site no.22, Figs. 5.3, 5.15). The avalanche had started in a tributary to Rakasjokk, flowed into the main stream and followed it for about 500 m downstream. It occurred in an area much frequented by tourists both winter and summer. The avalanche started on a slope of about $15-20^\circ$ gradient as a slab failure of the upper 20-30 cm of the snow cover, which had slid downslope on an underlying 30-70 cm deep snowpack. The size of the scar left by the avalanche was 6 m wide and 30 m long. Ram strength of the snow cover immediately upslope of the scar is shown in Fig. 5.16. The upper 30 cm had appreciable strength and correspond to the layer that failed and slid away. The lower 70 cm consisted of a very weak layer of depth hoar crystals. A basal ice crust was not present at the release scar. Further downslope the whole snow cover had been swept away by the avalanche, exposing the stream bed with a thin ice crust. Snow levees occurred on both sides of the track.

Kårsavagge

An avalanche from the same warm period in early June, 1982 occurred at a locality in Kårsavagge trough valley SW of Abisko. The site was along a locally steep stream course on the valley side south of lake Vuolemus Kårsavaggejaure (no.18 in Fig. 5.3). Field inspection was made on June 17 (Fig. 5.15). The avalanche had started at the upper rim of the steep slope (30°) by failure of the 1.5-2 m thick snowpack along vertical fracture planes,



Fig. 5.15a. Slush avalanche at a tributary to Rakasjokk near Björkliden. Occurrence early June, 1982. The major stream Rakasjokk runs diagonally from upper right to lower left in the picture. Note the wide flow track of the avalanche (more than 50 m), overflowing the small stream channel. The main part of the avalanche debris was deposited several hundred m downstream along the Rakasjokk stream. Photo 14.6.1982.



Fig. 5.15.b. Snow and slush avalanche at Kårsavagge, probably released in the beginning of June, 1982. Most of the snow consisted of clean, hard blocks, which were deposited especially on the left-hand part of the avalanche fan at the foot of the slope. Dirty, slushy snow continued for a longer distance to the distal, right-hand part of the fan. Photo 17.6.1982.

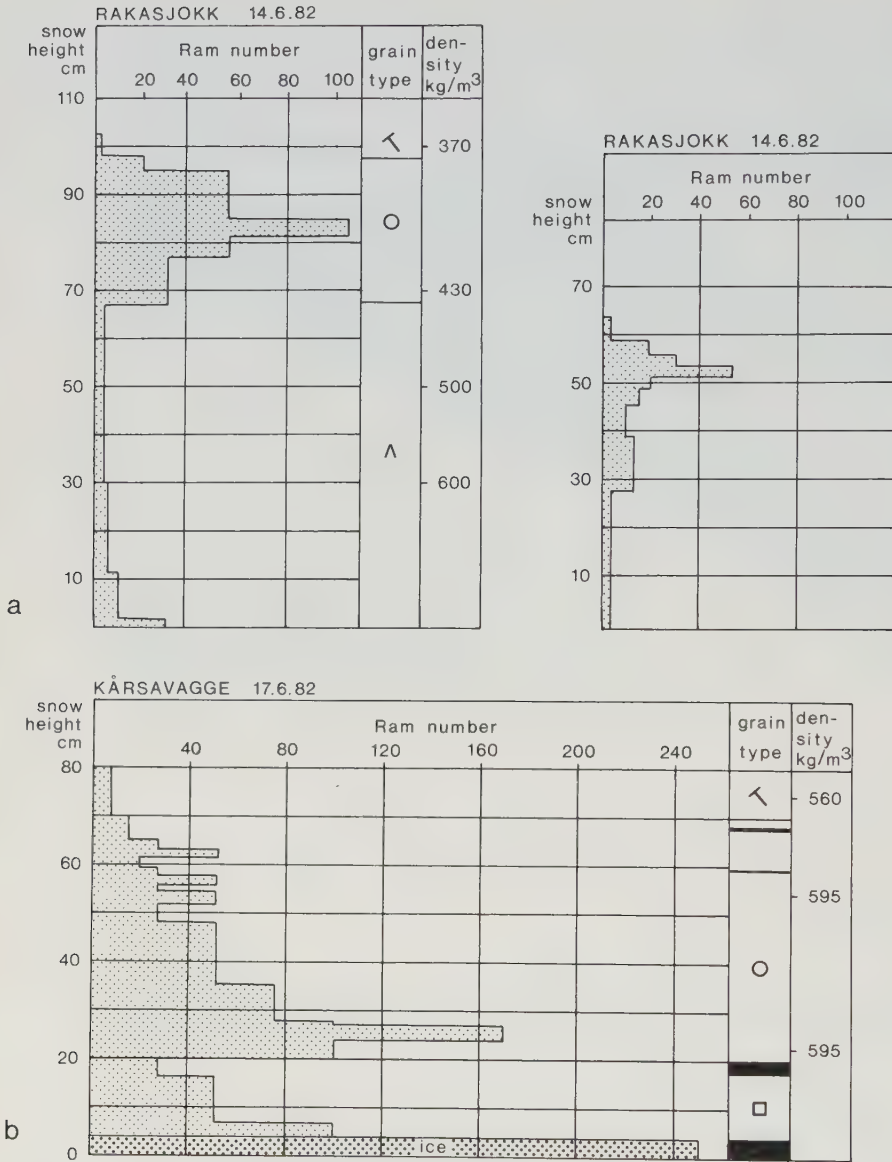


Fig. 5.16. Snow conditions in the release areas of two slush avalanches, which occurred during a warm period in early June 1982. a. Rakasjokk, b. Kårsavagge. The sites were investigated on June 14 and 17, 1982. Cf. text.

thus as a slab avalanche. Presumably, damming of meltwater had increased the weight of the wet snow, which rested on steeply inclined bedrock outcrops. Parts of the mass of snow slabs had been converted to slush as the stream broke up.

A few hundred metres upslope of the knickpoint where the avalanche started, minor breakups or slushflows were indicated by small snow levees and lobes along the stream course, similar in appearance to the snow cover at Mt. Njulla upslope of the waterfall at Ridonjira after the avalanche on June 2.

A Ram profile was measured in the stream bed snow cover, 30 m upslope of the release scar (Fig. 5.16). At the bottom, an ice crust, about 5 cm thick, rested on the ground. Between the hard, upper part of the snowpack and the weak lower TG-metamorphism layer was a thawed ice crust, which probably had served as a glide plane for the avalanche. Due to the steep slope at the point of the release scar, the actual conditions there could not be studied.

Nissunvagge

A final case to be reported was observed in Nissunvagge, occurrence in May 1984, exact date unknown. This avalanche came down the stream gully dissecting a glacifluvial terrace on the western valley side in the north part of Nissunvagge (Cf. Map I). The avalanche deposits in the valley bottom were investigated on May 29. The starting zone with a release scar 1.5 m high was situated approximately 300 m above the gully of the terrace, along a slope of 15-17° inclination. The avalanche had been gliding on top of a basal snow layer. Erosion of the ground occurred further downslope of the starting zone.

D. Concluding discussion on snow conditions

The internal changes affecting the snow cover at the study sites during the progression of the late winter to spring season can be summarized from the observations as:

1. Increasing density as old snow is gradually compacted and influenced by metamorphism. Typical values range between 300-600 kg/m³, usually increasing with snow depth.
2. Decreasing strength. Strength variations are initially enforced as weak layers stand out from strong refrozen ice-crusts after periodic melt-freeze cycles, but as melting continues they diminish in association with a general decrease in strength of the entire snowpack. A weak TG-layer appears to be common at the bottom.
3. Grain coarsening as a result of MF-metamorphism.
4. Gradual warming up to 0°C by refreezing of melt water and release of latent heat.
5. Increased amount of liquid water in the snowpack, opening up drainage paths along impermeable strata and as vertical drains. Slush is formed as thin layers above ice crusts, but probably only rarely as deeper accumulations due to impeded drainage. Liquid water contents probably seldom exceed 10-15% by volume in spring time snowpacks.

Two different modes of failure of the snow cover in stream courses seem likely as mechanisms leading to the release of slush avalanches:

1. Water-saturation release

This mechanism involves the saturation of the snow cover by water, thereby converting it into slush. The extremely low strength of snow slush allows it to start moving also on very low slope gradients. Movement may start at a point of intersection of the snow surface and a watertable within the snow, and be transmitted upslope by headward growth (cf. Nobles 1966). It may be due to either of two triggering causes: a) some sudden input of further water, e.g. after breaching of dams formed by collapsed snow banks upslope, to mobilize the accumulated slush, or b) bursting of a barrier downstream holding the slush back. During the study period, no certain example of this type of initiation was observed, although the cases at Njulla and Kårsavagge in 1982 had indications of at least a partial influence by this mechanism.

2. Snow slab release

The mechanism is analogous to the release of snow slab avalanches in its initial stage, followed by the transformation of the sliding snow mass into slush during its descent downslope along the stream course, together with large amounts of released melt water. The slush formation may also take place with a time lag, where the avalanched snow masses cause damming of a stream. The slab release occurs e.g. in steep frozen waterfalls (case Kaise-pakte 1979). The failure of the snow and ice masses can be caused by loss of strength and increase in weight due to melting. In other cases, the failure of the snow slab probably differs from the normally observed for snow avalanches, as it occurs also on slopes less steep than 20° . A possible reason for this may be the influence of a 'hydropressure' effect (Akkouratov 1976) caused by subnival stream flow, or concentrated flow along certain snow strata, due to rapid melting. This may be augmented by loss of strength along the horizons affected, e.g. above impermeable ice crusts. Such crusts are associated with water retention and may act as sliding surfaces for slush avalanches (Cottman 1966).

Of the observed cases the slab release mechanism seems plausible for Njulla, Kårsavagge, Rakasjokk and Nissunvagge. The slide plane has either been on top of a weak layer of TG-metamorphism grains (depth hoar), or on a basal ice crust covering the stream bed. The former type is common also for winter snow avalanches, whereas the latter is more specifically related to avalanche occurrence on stream bed sites (cf. Rapp 1960b).

Formation of slush above ice layers in snowpacks has been noted from many areas (e.g. Scotland; Ward 1980). Basal ice layers on stream beds are furthermore reported as widespread in high Arctic areas during the melt period (cf. Woo et al. 1982), and snow jams frequently interfere with the normal runoff pattern (Woo & Sauriol 1981). The observations reported here indicate similar conditions in the mountains of northern Scandinavia. The importance of ice crusts with respect to both melt water retention in the snowpack, and as glide planes for avalanches, suggests a possible relationship to slush avalanche occurrence. However, the difficulties of making observations of the conditions at starting

zones shortly after the avalanches occurred, due to their inaccessibility, have limited the number of studied cases so far. Different sites appear to be activated in response to strictly local conditions, although in some years more general release over large areas may occur, as in 1982, when late spring avalanches occurred not only in Swedish but also in Finnish Lapland (Matti Seppälä, pers.comm.).

5.3.6. Summary and conclusions

The release of slush avalanches constitutes a special kind of stream breakup active in years when the development of drainage paths in the snow cover of stream channels is insufficient to deal with the meltwater discharge in spring. This imbalance may be due to either unusual conditions in the snow cover causing inefficient drainage, or exceptionally large momentary inputs of melt water after sudden rises in temperature.

The observations from the Abisko area indicate a critical period for slush avalanche release a few days after marked warming sets in. Continued high temperatures (several degrees above zero) once efficient drainage of the snow cover has been established will not lead to slush avalanche occurrence, with the possible exception of situations of interference of the drainage network capacity by snowbridge collapses, creating snowjams.

The incidence of the critical periods varies between different years, but is normally confined to the period middle of May to early June in the Abisko area.

5.4 Site characteristics of slush avalanches

5.4.1 Introduction

In the course of the survey of the geomorphological impact of avalanches along stream courses in the study area, qualitative and quantitative data on site characteristics were collected. Thus a more comprehensive treatment of this topic can be given for slush avalanches than was possible for debris flows (cf. ch. 4.2.4, 4.3).

5.4.2 General characteristics

The following site attributes and variables were included in the survey, partly following Lied & Bakkehöi (1980):

1. Latitude
2. Longitude
3. Estimated altitude of starting zone
4. Altitude of runout zone (avalanche debris deposit)
5. Estimated height of avalanche path
6. Gradient of starting zone
7. Gradient of lower track upslope of runout zone
8. Mean path gradient, from the upper end of the starting zone to the lower end of the runout zone

9. Aspect, classified into 8 sectors of 45 degrees
10. Longitudinal profile of the path
11. Cross profile of the track
12. Existence of a distinct topographic step in the starting zone
13. Starting zone morphology
14. Runout zone morphology
15. Type of avalanche deposit
16. Existence of lake upstream of starting zone
17. Stream order
18. Runout zone situated in forest
19. Bedrock

Values for variables 6 to 8 were measured photogrammetrically, as described in chapter 3, for a limited number of paths. Terrain profile type was only determined for these sites. It should be noted that the included data are of variable accuracy. A certain degree of subjectivity is inevitable where chosen classes may show gradual transitions in their characteristics. This should be kept in mind when judging conclusions based on the statistic treatment of the data, although mainly non-parametric tests have been applied due to the nature of the data.

Results and discussion

A. Altitudinal distribution

The altitudinal distribution of 224 sites is illustrated in Figure 5.17. Values both for altitude of starting zone and runout zone are approximately normally distributed. In general, as shown by the means, the sites occur entirely above the forest limit (at about 500-700 m a.s.l. in the area). 17% of the paths (39 out of 224) reach into forested areas with their deposits. All starting zones occur above the forest limit, the highest situated at an altitude of about 1500 m a.s.l.

B. Estimated height of paths

An avalanche path includes starting zone, track and runout zone (Perla & Martinelli 1975). The mean height of the paths, that is, the vertical displacement of the avalanches from starting zone to runout zone, is about 330 m (224 sites). The range of values is great, from merely 40-50 m up to about 1000 m height difference. Two examples of paths with extremely low vertical displacement are shown in Figs. 5.18 and 5.19 from the area E of Mt. Låktatjåkka. The height is usually somewhat larger for paths that begin at the higher altitudes. Maximum reported values for snow avalanche sites in more alpine areas than Swedish Lappland are considerably higher, exceeding 1500 m. For instance, in the mountains of Montana, USA, the mean height of 440 avalanche paths was about 600 m (Butler 1979). However, note that the present survey refers to slush avalanches, occurring along stream courses. The mean height difference for snow avalanche sites, usually located to steep mountain slopes, is larger also in Lappland. Some quantitative data verifying this statement are given in section 5.4.3.

C. Gradients of paths

Data on slope gradients of 75 stream courses followed by avalanches are summarized in Figure 5.20. Earlier descriptions of slush avalanches have shown the extremely low values of slope gradient for this avalanche type compared with normal snow avalanches (e.g. Nobles 1966).

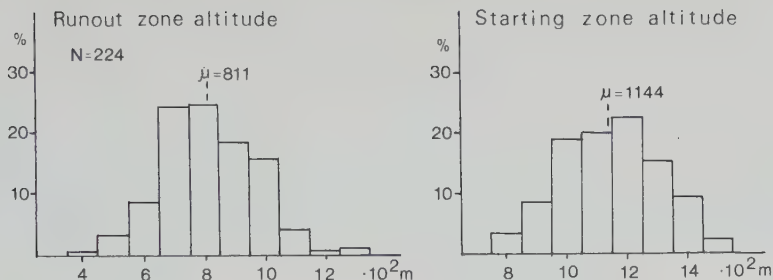


Fig. 5.17. Altitudinal distribution of slush avalanche sites in northern Swedish Lappland. Most sites occur entirely above the forest limit at about 500-700 m altitude.

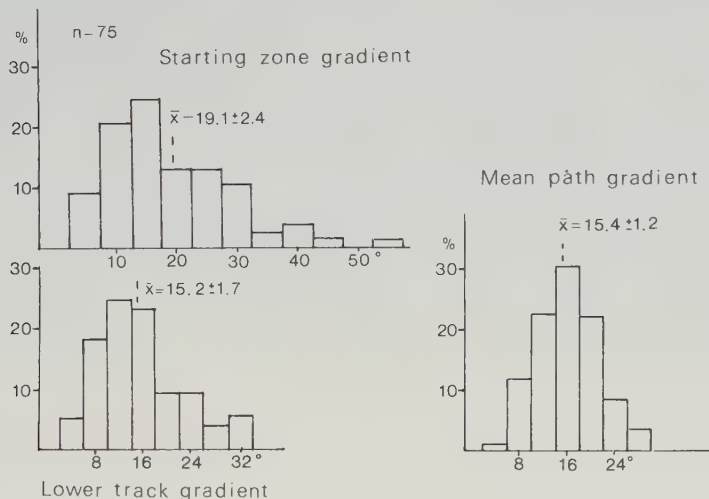


Fig. 5.20. Slope gradients of slush avalanche paths in northern Swedish Lappland. Based on sample of 75 paths. Measurements made photogrammetrically on vertical air photos. Confidence limits given at the 5% significance level.

The distributions of the measured slope angles are right-skewed, or approaching normal in the case of the path gradient. The majority of these slope gradients refer to first order streams. Mean channel slopes for streams of a given order generally exhibit a log-normal (right-skewed) distribution according to Chorley (1966). The values found in the present study are compared with some data on snow avalanches in northern Lappland and other areas in Table 5.3.

Although some mixed snow/slush avalanche sites may be included in the measured data set, the mean value is very low compared with the values given for pure snow avalanche sites. Even the lowest individual gradient found in the Norwegian study (18°) is higher than the mean of the present study, which strongly suggests that the data largely refer to different types of avalanches. This is further underlined by the mean gradient of the starting zone for the Lappland sites (19°) which is lower than what is normally considered as necessary for the release of snow slab avalanches (about 25° , e.g. Perla 1980). The lowest value measured was 6° .

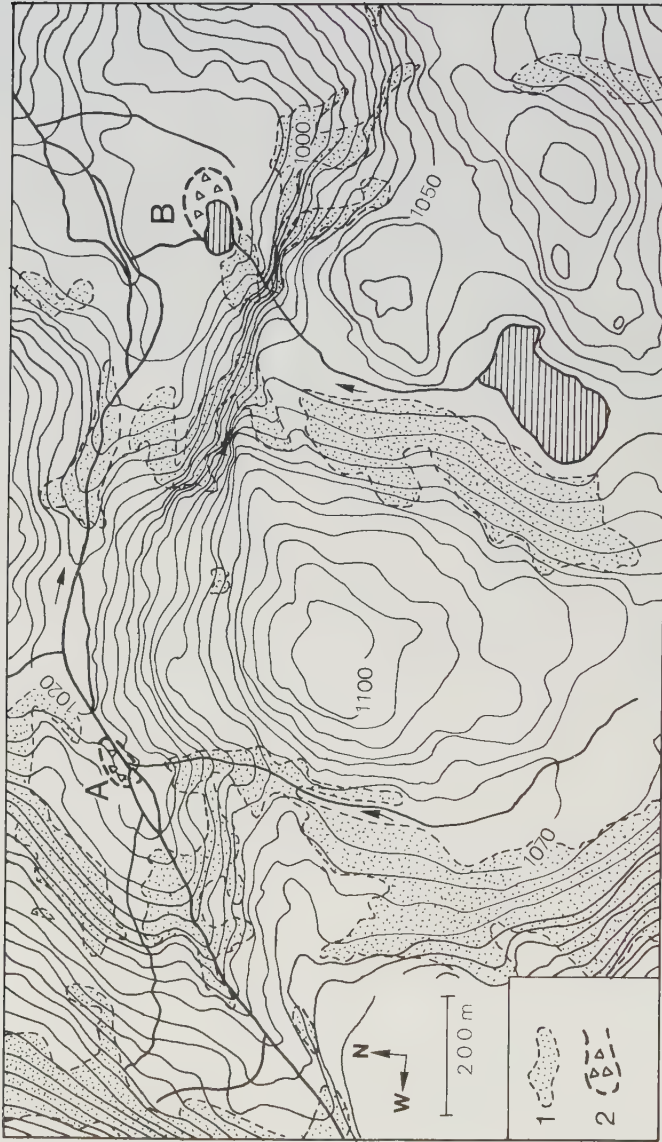


Fig. 5.18. Two slush avalanche paths along streams in the area east of Mt. Låktatjåkka, western Abisko mountains. Key: 1. long-lasting or perennial snow field, 2. debris coating in runout zone of slush avalanches. Both avalanche paths are notable for their extremely low vertical height, about 40-50 m. Site A was described by Rapp (1960b). The sites provide examples of slab release of thick snowdrifts accumulated on bedrock thresholds in convex reaches of stream courses. Contour interval 5 m.



Fig. 5.19. Snow and slush avalanche site W of Mt. Låktatjåkka (A of Fig. 5.18). Observations in 1984 and 1985 have given the following information on the behaviour of the site regarding avalanche release: The snow patch in the middle of the picture attains a thickness of 5-10 m each winter, covering also the bedrock threshold in the stream course to the left. During melting in the spring, the snow cover, resting on the bedrock, which is partly ice-covered, breaks up along vertical fractures and a snow slab avalanche is released. In spite of the low vertical drop (30-40 m), the avalanche may continue for 50 m upslope on the opposite side of the main stream, depositing blocks of 0.5 m size (foreground). Parts of the avalanche snow are converted to slush in the main stream and continue downstream for several hundred m. Photo August 1982.

The inclination of the avalanche track immediately above the runout zone was included mainly to see if any relationship existed between this variable and the type of deposit or particular impact features in the runout zone. A steep angle and a sharp break of slope would imply a sudden retardation of the avalanches through impact pressures, whereas a gentle angle would permit a more gradual slowing down and a longer horizontal reach of the avalanches. The mean inclination of this lower part of the track was 15° , however with a very large range of values, from 3° to 33° .

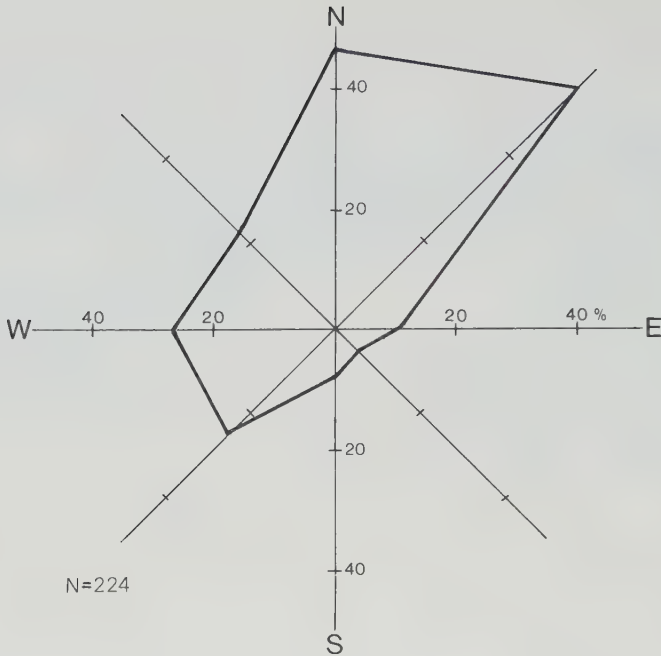


Fig. 5.21. Aspect of slush avalanche sites in northern Swedish Lapland.

D. Aspect

The aspect (orientation) of the sites, particularly with respect to the starting zones, was determined from topographic maps. Only a rough classification into eight sectors of 45° was considered as justifiable, though still sufficient to reveal any marked preferred orientations.

The differential distribution of snow cover in the terrain could be expected to influence the location of sites, in a similar manner to its influence on the distribution of glacial cirques and recent glaciers, favouring E- and NE-facing slopes (Enquist 1916, Vilborg 1977, Östrem et al. 1973). Shaded and leeward slopes generally have the deepest snowpacks, which makes them particularly subject to snow avalanches, as is well known from many areas (e.g. Gardner 1970). If, on the other hand, insolation is an important factor in the release mechanism of slush avalanches, through its effect on snow melting, one would expect a high percentage of sites with a southwesterly aspect.

An inspection of the orientation diagram of Figure 5.21 reveals a concentration of sites in the northern and north-eastern sectors (51% together). However, sites facing westerly directions are also fairly common. Notable is the comparative lack of sites with aspects from E to S (only 12% together).

The distribution of aspect for the investigated sites may indicate a complex interaction between snow distribution in the terrain and slope orientation with respect to insolation as a

Table 5.3 Slope gradient data for avalanche paths (degrees)

area	mean path gradient	range	no.of sites	reference
N.Lapland				
(snow aval.)	30.3	21-36	30	this study
(slush aval.)	15.4	6-29	75	" "
Norway	33	18-50	423	Lied & Bakkehoi (1980)
Colorado	26		67	Bovis & Mears (1976)
various mts, USA	38		194	Perla (1980)

Table 5.4. Terrain profile characteristics of 75 avalanche paths along stream courses

long profile		frequency		
convex/concave		concave	convex	stepped
	11	30	9	25
cross profile				
	unconfined	moderate incision	strong incision	
	9	33	33	
(chi-square test: sign. 1 %)				

control of geomorphologically significant slush avalanche activity. The fairly common westerly aspect may also be due to the influence of a higher accumulation of wet snow on windward slopes, which could counteract the normal wind redistribution of snow toward easterly slopes. The relationship of aspect to other studied site parameters will be dealt with in section 5.4.3.

E. Terrain profile

The longitudinal profile of 75 paths was classified into dominantly concave, convex, convex-concave (upper slope section convex, lower concave) and stepped, using information from the topographic map and from air photo stereo pairs.

Concave profiles were found to be more common than convex ones, or combined convex-concave (Table 5.4). A high percentage of paths display a stepped profile due to bedrock structure. A marked topographic step or threshold occurs in the starting zone area in 38% of all paths surveyed.

Cross profiles of the 75 paths were classified to estimate the degree of incision of the stream courses, thus whether the avalanches were of the 'channelled' or 'unconfined' type, and compared with a random sample of unaffected stream courses (see section 5.4.3.).

The majority of paths follow moderately or strongly incised stream channels (Table 5.4). Detailed terrain profiles based on field measurements of a number of type cases are presented in section 5.5.

F. Morphology of starting and runout zones

The type of morphology of the starting zones could influence the disposition towards slush avalanching as a stream break up method in spring at various sites. Therefore, a generalised classification of starting zone morphology was applied to the surveyed sites to detect possible preferred situations. The result is presented in Table 5.5, where data are given for 75 sites (sample identical to that used for collecting slope gradient data). The sample has also been stratified into avalanche sites with starting zone inclination below and above 20° , to note possible contrasts between gentle sites of exclusive slush avalanche origin, and steep sites where snow avalanching may also occur.

Three types of starting zones are especially common, namely glacial cirques, shallow concave depressions and incised scars in bedrock. Less common are flat and convex slopes, which are more typical of snow avalanche sites (Lied & Bakkehöi 1980). Looking at steep starting zones separately, these seem to be particularly associated with scars and flat slopes. Glacial cirques are poorly represented among the steep starting zones, probably because the release area generally seems to be situated at the threshold, which is seldom very steep. However, the sample size is too small for more general conclusions to be drawn.

Runout zone morphology for surveyed cases is shown in Table 5.5. A stratification has been made according to the inclination of the slope segment immediately upslope of the runout zone. The majority of tracks are inclined less than 20° in this part.

Table 5.5. Morphologic classification of starting and runout zones of 75 avalanche paths along stream courses

<u>starting zone</u>	frequency			
	all sites	gradient < 20°	>20°	
cirque	18	13	5	
depression	21	12	9	
scar	21	9	12	
flat slope	9	3	6	
convex slope	6	5	1	
(chi-square test: all sites sign. 5 %, split sample non-sign.)				
<u>runout zone</u>	all sites	<10°	11-20°	>20°
valley bottom or other flat ground	41	16	16	9
V-shaped valley	11	2	8	1
topographic obstacle	5	1	3	1
lake	11	2	4	5
open slope	7	3	3	1
(chi-square test: all sites sign. 5 %, split samples non-sign.)				

G. Type of deposit

Avalanche deposits along stream courses were classified into four groups, based on morphological characteristics and terrain position: 1. slush avalanche fans, 2. multiprocess cones (slush avalanche/alluvial cones), 3. slush avalanche boulder tongues and 4. impact pools. Two special varieties were distinguished for type 3: a. stream-crossing tongues and b. tongues with impact pit. Descriptions of the different types will be given in section 4.5. The most common type of deposit due to slush avalanches along stream courses in the study area is no. 3, followed by the multi-process cones (type 2), Table 5.6. About 10 % of the avalanche paths lack well-defined terminal deposit (type 5).

H. Hydrography

Mainly small first-order streams are subject to slush avalanching in northern Lapland, constituting 78% of affected streams. The hydrological effects, as a high-level discharge through release of ponded meltwater following the avalanche, will be noticeable further downstream in second- and third- order streams of a drainage basin. Increased sediment yield in the streams will probably also be propagated for considerable distance downstream from the actual runout zone of the avalanches. One instance where this has been recorded is described by Rehn et al. (1982) from the Kärkevagge basin.

The presence of a lake upstream of a starting zone could have an influence on avalanche frequency. A rise in lake water level during the snow and ice melting, coupled with snow damming of the lake outlet, might constitute a possible triggering condition for the avalanches. Blocking of outflow channels by thick snowdrifts and catastrophic bursting of the snow jams releasing a peak runoff is known to occur frequently in arctic ponds or small lakes (Woo 1980, Woo & Sauriol 1981). The frequency of lakes present upstream of a starting zone for slush avalanches in the investigated area was 23%. See further section 5.4.5.

I. Bedrock

A uniform frequency of avalanches in an area with variable bedrock might have a much stronger geomorphic impact on certain, weaker types of rock than on hard, resistant rocks. However, since different types of bedrock are associated with different relief conditions, an apparent association of type with geomorphic impact may also be related to this factor.

Table 5.6. Morphologic classification of avalanche debris deposits along stream courses (N=224)

	frequency
1. Slush avalanche fans	9
2. Multiprocess cones	53
3. Slush avalanche boulder tongues	89
a. " across streams	29
b. " with impact pit	11
4. Impact pools	10
5. Indistinct form	23

Within the study area, 34% of the avalanche sites occur in areas characterized by various schists, 24% in dominantly amphibolite areas. A fairly large group of sites (21%) have their starting zones in amphibolitic bedrock and their runout zones on schists. A smaller percentage occurs on granitic rocks or exposed basement rocks (8% each). Further notes on bedrock influence on avalanche sites are given in the following section.

5.4.3 Relationships between site factors

A. Differences in site gradient, altitude and relative height

Differences in site factors between the categories distinguished on morphological grounds are shown in Table 5.7. The relative height of the paths was lower for impact pools and pits and slush avalanche fans compared with boulder tongues. Paths with gentle starting zone gradients (less than 20°) were generally lower, that is, the avalanches had a smaller vertical displacement than steeper paths. Mean path gradient for different categories deviated 2-3 degrees from the mean for all paths. The slush avalanche fans and the multiprocess cones had paths with gentle gradients, while paths ending in a boulder tongue or impact pool were steeper. Large variation was found in the means for starting zone gradient. Fan- and cone-type of deposits seem to be associated with sites, where the avalanches are released on fairly gentle slopes, in contrast to the boulder tongues. No marked variation in means were found for lower track gradient, hence its possible influence on the morphology of the deposit, or other features such as impact forms, appears to be of limited importance.

A Kruskal-Wallis analysis of variance by ranks indicated that the measured differences are very unlikely the result of chance (significance level 1%), but represent actual variations between the morphological categories, assuming no bias in the sample.

Table 5.8 presents results on the differences in starting zone altitude and path length related to morphological factors. The distributions show for instance that flat and convex slopes as starting zones are most common at low altitude and associated with avalanche tracks of comparatively short length. Scars as starting zones make up the other extreme. Long avalanche paths are very significantly linked with incised stream channels.

Table 5.7. Numerical data for avalanche paths along stream courses according to the type of deposit.

type of deposit	height of path		altitude of starting zone		altitude of deposit		starting zone gradient		mean path gradient		track gradient above runout zone	
	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s	$\bar{x}$	s
FANS	283	126	1077	163	794	100	15.4	6.2	13.0	2.9	16.4	8.1
MULTIPROCESS CONES	349	182	1197	127	848	157	12.3	6.1	13.1	5.2	14.6	6.8
TONGUES	450	186	1210	194	760	165	23.8	10.2	16.6	4.6	15.1	8.3
STREAM-CROSSING TONGUES	359	173	1165	138	806	121	20.9	11.5	16.1	6.1	13.8	6.8
TONGUES WITH IMPACT PITS	255	93	1261	202	1006	210	19.0	12.1	12.9	4.5	15.0	8.4
IMPACT POOLS	277	37	938	81	662	94	15.3	7.9	18.5	4.6	18.0	6.1
ALL PATHS	358	168	1174	177	807	167	19.1	10.3	15.4	5.1	15.2	7.4

B. Relationships with respect to aspect

An interesting relationship was found between aspect and altitude of the starting zone (Table 5.8). The most common aspect, N-NE, proved to be associated with fairly low starting zone altitudes, while zones facing SW-W-NW were mainly located to higher altitudes. One reason may be that the predominant snow drifting towards easterly directions supplies N-NE-facing slopes with more snow down to lower altitudes, compared with the slopes with westerly aspect. Possibly also the higher insolation on westfacing slopes causes some melting at lower altitudes before the general rapid spring warming sets in, thus leading to a lower probability of slush avalanches in these locations.

There appears to be a slight difference in steepness of starting zone for the two aspects, with steeper values for N-NE-facing sites. This could possibly be related to bedrock structures, which tend to favour steep walls with easterly aspects.

C. Relationships with respect to bedrock

Avalanche paths along streams in the dominating bedrock types of the area were compared as to bedrock-related differences in site factors (Table 5.9). The high relief of the amphibolite massifs mentioned previously is reflected in significantly higher avalanche paths (U-test). Differences between the other bedrock types are not significant.

No difference in mean gradient was found for amphibolite vs. schist sites (U-test). Concerning possible structural influence on site gradients, see above regarding aspect of starting zones.

The relationship of morphological characteristics of the sites to bedrock is indicated in the Table, using frequency as a criterion of bedrock influence. Due to the low number of cases for certain combinations, no firm conclusions can be drawn, although some features of the frequency distribution seem to support field observations, e.g. the tendency towards concave profiles on amphibolite slopes vs. stepped profiles on schists.

D. Lengths of paths

The importance of terrain factors for the paths lengths or total runout distances of avalanches has been investigated for predictive purposes in several areas (Bovis & Mears 1976, Lied & Bakkehöi 1980, Fitzharris 1985). In the present study, the lower end of the debris accumulations was used as a morphological criterion for maximum runout distance or path length. This may not include extremely rare large avalanches, the morphological signs of which could be obliterated or weakened in long time spans.

A significant correlation ($r=0.57$) with path length was found for starting zone altitude (Fig. 5.22), perhaps indicating deeper snowpacks and larger avalanches at high compared with low levels.

Table 5.8. Relationships between site factors at slush avalanche paths

	Starting zone altitude by morphology, m a.s.l.			tot
	700-1100	1100-1300	1300-1500	
cirque	3	10	6	19
depression	10	10	3	23
scar	3	8	8	19
flat/convex slope	10	4	2	16

(chi-square test: sign. 5 %)

	path length by degree of incision, m				tot.
	800	800-1200	1200-2000	2000	
none or moderate	11	18	12	3	44
strong	2	8	12	10	32

(chi-square test: sign. 1 %)

	Path length by starting zone morphology, m			tot.
	1000	1000-1500	1500	
cirque	4	9	5	18
depression	12	7	3	22
scar	5	7	9	21
flat/convex slope	10	4	2	16

(chi-square test: sign. 5 %)

aspect	Starting zone altitude by aspect, m a.s.l.			tot
	700-1000	1000-1200	1200-1500	
N-NE	12	19	14	45
SW-W-NW	1	9	17	27

(chi-square test: sign. 1 %)

aspect	Starting zone gradient by aspect			tot.
	14°	15-28°	28°	
N-NE	14	20	11	45
SW-W-NW	14	11	2	27

(chi-square test: non-sign.)

Null hypothesis of no differences in frequency distributions tested at the rejection level of 5 %.

Table 5.9. Variation in site factors due to bedrock

	amphibolite N=54	schist N=77	granite N=19	basement	comment
path height, m	μ 394 σ 161	μ 275 σ 139	μ 310 σ 168	μ 234 σ 92	
mean path gradient	n=31 13.6	n=35 15.1			non-sign. difference (u test)
path incision	frequency				
none or moderate	21	22	6		non-sign.
strong	20	10	3		(chi-square test)
path long profile					
convex/concave	5	4	3		non-sign. (chi-square not applicable)
concave	21	8	3		
convex	3	3	-		
stepped	12	15	3		
starting zone morphology					
cirque	12	5	2		"
depression	12	13	3		
scar	11	5	2		
flat slope	4	4	1		
convex slope	2	3	1		

Gradient of the paths are, surprisingly to some perhaps, negatively correlated with length. An explanation might be that for steep paths avalanches on the average are smaller, released from thinner snowpacks, than for more gentle paths, where more snow may accumulate before avalanching occurs. A similar relationship has been found for snow avalanches (Lied & Bakkehöi 1980). Also the ability of slush to move even on very low gradients may be evoked to explain the long range on gentle slopes.

The presence of lakes upstream of slush avalanche starting zones tends to be associated with topographic thresholds. These factors appear to be linked with short paths, perhaps indicating an effect favourable to more frequent release of small avalanches. However, the presumably very complex multivariate nature of the interactions between terrain factors and avalanche size makes this assumption speculative for the present.

The established site factor relationships are summarized in Fig. 5.23, which shows the associations between factors, with an indication of the nature of the relationships (positive or negative correlation) where applicable.

5.4.4 Comparisons with snow avalanche sites

To highlight the distinguishing characteristics of slush avalanche sites, comparative quantitative data were sampled for 30 snow avalanche paths, chosen at random within the available air photo coverage. The paths were identified on the basis of basal debris accumulations (avalanche boulder tongues). They occur on mountain slopes and are not associated with any permanent stream courses. The comparison was limited to 7 variables, extracted from topographic maps. The sampled data are summarized and compared with data for slush avalanche paths in Table 5.10.

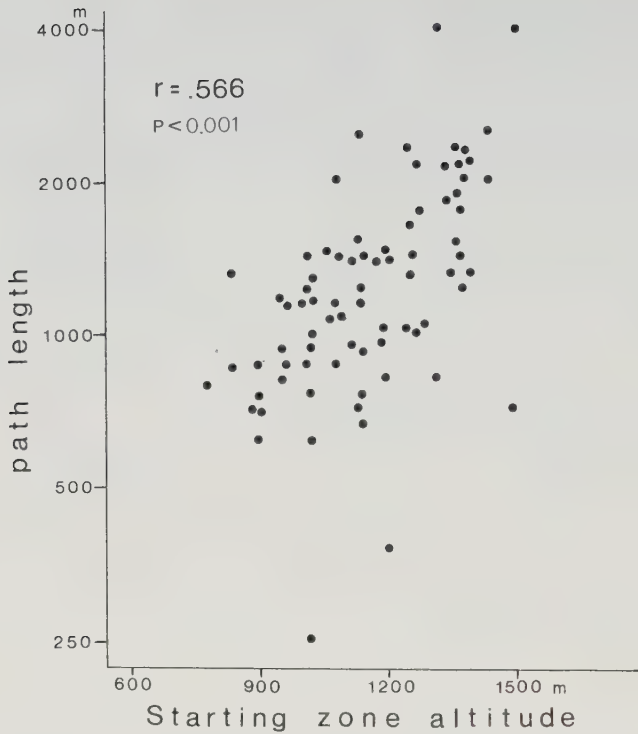


Fig. 5.22. Relationship between path length and starting zone altitude for 77 slush avalanche sites in northern Swedish Lapland. Paths lengths measured on maps and air photos on the basis of morphological signs (debris accumulations). Starting zone altitude estimated on topographical maps, using morphological information from air photos.

The comparison shows that slush avalanche paths are significantly flatter, tend to occur at lower altitudes and are generally lower in relative height than snow avalanche sites. Despite the lower height and less steep path, the horizontal length is much larger for slush avalanches, indicating either the different flow properties of water-saturated vs. dry or only slightly wet snow, or a generally larger size and momentum of slush avalanches. Whereas the snow avalanche sites of the sample all lacked a topographic threshold in the starting zone, more than one third of the slush avalanche sites had distinct thresholds.

The aspect of the snow avalanche sites showed a higher concentration to easterly and southerly orientations than for slush avalanches. Although the distribution is significant, it is likely that a larger sample is needed to obtain a representative view of the aspect factor for snow avalanches. However, data compiled by Schmacke (1959) also indicate a preponderance of easterly directions over northerly for snow avalanches in Lapland.

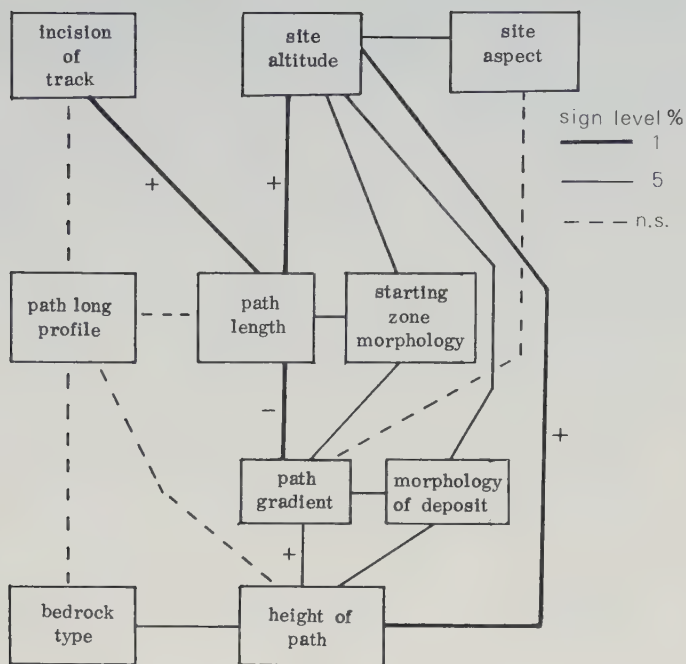


Fig. 5.23. Relationships between site factors at slush avalanche paths in the study area. Positive or negative association shown where applicable. n.s. = possible relationship, statistically non-significant at the 5 % level. Based on data in Tables 5.5-5.10.

5.4.5 Comparisons with non-avalanche stream courses

To obtain some information on what makes certain stream courses subject to slush avalanching with geomorphic effects, a number of unaffected stream courses, chosen at random, were also studied concerning slope gradients and morphology. A calculation of the total number of first-order streams in the study area showed that about 5% were affected in a geomorphologically important way by slush avalanches. Two gradient variables were measured photogrammetrically in air photos. The time-consuming work involved in this task limited the size of the sample to 20, which may be considered on the low side for meaningful comparisons. However, as distribution-free statistics were used, also small samples can be analyzed. The result of the comparison is given in Table 5.11.

With regard to slope gradients, the samples indicated that stream courses affected by slush avalanches are slightly steeper than the majority of first-order streams and have a predominant N-NE aspect. The sample of non-avalanche stream courses showed an aspect distribution that did not significantly differ from one resulting by chance. No particular terrain features seem to distinguish slush avalanche affected stream courses from other

Table 5.10. Comparison of data for snow avalanche and slush avalanche sites

	snow aval. n=30	slush aval. n=77	Mann-Whitney U-test sign.level
mean path gradient (degr.)			
mean	30.3	15.7	1%
median	31.0	15.0	
st.dev.	3.7	4.9	
min.	21.0	6.0	
max.	36.0	29.0	
starting zone altitude (m)			
mean	1556	1166	1%
median	1565	1160	
st.dev.	135	176	
min.	1160	760	
max.	1860	1500	
runout zone altitude (m)			
mean	1142	807	1%
median	1177	790	
st.dev.	132	167	
min.	780	460	
max.	1370	1300	
path height (m)			
mean	429	358	5%
median	420	320	
st.dev.	151	168	
min.	160	50	
max.	710	980	
horizontal length (m)			
mean	741	1339	1%
median	706	1194	
st.dev.	315	676	
min.	155	360	
max.	1600	4070	
aspect (frequency %)			
N	-	31.2	chi-square test sign.level
NE	10.0	29.9	
E	16.7	2.6	5% snow a.
SE	33.3	1.3	
S	13.3	-	
SW	13.3	9.1	
W	-	16.9	
NW	13.3	9.1	<1% slush a.
topographic threshold			
absent	30	47	
present	0	30	

courses, except degree of incision, which is higher for the avalanche sites. Whether this is a result of long periods of avalanche erosion, or simply implies an increased probability of slush avalanche occurrence in fluvially incised streams due to favourable conditions for snow accumulation, cannot be concluded from the data.

Table 5.11. Comparison of data for stream courses with and without geomorphic impact of slush avalanches

	frequency		chi-square test	
	with n=77	without n=20	sign.level	
			<u>slush a.</u>	<u>streams</u>
'starting zone' morphology				
cirque	19	7		
shallow depression	23	7		
scar	19	3		
flat or convex slope	16	3		
			non-sign.	
aspect (%)				
N	31.2	5		
NE	29.9	20		
E	2.6	20		
SE	1.3	15		
S	-	10	5%	non-sign.
SW	9.1	5		
W	16.9	25		
NW	9.1	-		
		(N=888)		
lake absent (%)	83	69.5		
topographic threshold				
absent	47	9		
present	30	11		
			non-sign.	
terrain long profile				
convex, convex/concave	20	2		
concave	32	10		
stepped	25	8	non-sign.	5%
terrain incision				
none	10	6		
moderate	34	11		
strong	33	3	1%	non-sign.
bedrock				
amphibolite	29	6		
schist	24	7		
granite a.o.	24	7		
			non-sign.	
			Mann-Whitney U-test	
			sign-level	
'starting zone' gradient				
mean	19.4	15.6		
median	17.0	12.5		
st.dev.	10.1	10.6		
min.	6.0	2.0		
max.	53.0	37.0		
			non-sign.	
mean 'path' gradient				
mean	15.7	11.8		
median	15.0	11.5		
st.dev.	4.9	6.2		
min.	6.0	2.0		
max.	29.0	21.0	5%	

The frequency of lakes upstream of starting zones for slush avalanches was compared to lake frequency along all first-order streams in 11 third- to fifth-order drainage basins, comprising about 3300 km² of area. It turned out that 30% of all first order streams were associated with lakes, but only 17% of the streams with signs of geomorphic activity by slush avalanches. As mentioned, for the whole study area and including higher order streams,

a somewhat higher percentage (23) is obtained in the latter case. However, the data show that there is no preferred localization of slush avalanches to stream courses with lakes in the headwater areas. This does not exclude that differences exist in e.g. avalanche size or frequency between sites associated with lakes compared with other sites.

5.4.6 Summary

Site characteristics are described for localities (stream courses) with marked geomorphological influence of slush avalanches and compared with data for snow avalanche sites and first-order stream courses lacking avalanche influence. Distinguishing features for slush avalanche sites are fairly gentle gradients (mean 15.4°) and low vertical range (mean 358 m), but great lengths (mean 1.3 km) compared with snow avalanche sites. Longitudinal profiles are frequently stepped in their upper parts. Most sites face N or NE, less frequently W.

The starting zones are commonly glacial cirques, concave depressions or v-shaped scars, thus topographic positions favourable to snow accumulation and functioning as headwater areas for first-order streams. Compared with stream courses in general of this order, sites affected by slush avalanches tend to be slightly steeper and with higher degree of incision.

The morphology of debris deposits at slush avalanche sites appears to be related to the steepness and height of the path, with gentle and low paths associated with fan- or cone-shaped deposits, steep and high paths with tongue-shaped deposits. Observed relationships are summarized in Fig. 5.23.

5.5 Geomorphological effects of slush avalanches

5.5.1 Erosion and transfer of debris

Earlier observations have established the prevailing view of slush avalanches as geomorphic agents of high capacity (e.g. Rapp 1985). In the present study, two lines of evidence were explored to add further information on the erosional importance of the process:

1. estimates or measurements of accomplished erosion/deposition for recently observed cases.
2. estimates of thickness and extent of postglacially accumulated debris deposits.

The total geomorphic effect is dependent on frequency as well as magnitude of the process events. The frequency of slush avalanches will be treated in section 5.6. In this section, attention is focussed on the magnitude of erosion by slush avalanches, which will be discussed using evidence of the types mentioned above.

A. Estimates of recent erosion

The geomorphic capacity of individual snow avalanches is often less than 1 m^3 of transported rock debris (Rapp 1960b). Gardner (1970) measured quantities of 2 m^3 , 14 m^3 and 36 m^3 rock debris for three wet snow avalanches in Canada. Most of this material is not eroded directly from the bedrock by the avalanches, which merely sweep out weathered debris along their tracks. They may contribute to erosion of rockwall chutes (Rapp 1959, Markgren 1964) or influence slope gradient on steep walls (Rapp 1960b p.136, Hewitt 1972 p.33).

By contrast, for slush avalanches, debris loads of $200\text{-}300 \text{ m}^3$ have been reported (Rapp 1960b). However, as cautioned by Gardner (1983) the erosive effect of slush avalanches can be low or none, depending on local circumstances. Basal ice crusts on stream beds during the melt season in Arctic areas (cf. Woo et al. 1982) may prevent erosion. On the other hand, slush avalanches generally occur late in the spring and may move on to low snowfree footslopes, where the ground is more exposed to erosion, compared with the situation for most snow avalanches.

Measurements of rock and soil debris brought down by slush avalanches were made for two cases, Njulla 1982 and Nissunvagge 1982. The area occupied by the deposits was calculated from several transect tape measurements, and the volume estimated using a value for average debris thickness based on numerous checks along the transects. Where the debris was largely made up of larger clasts, these were counted in limited sections of the deposits and total number and corresponding volume assessed for the whole deposit. More rough estimates of debris volumes were made for several of the other observed avalanches.

The slush avalanche at Mt. Njulla June 2, 1982 brought down roughly 300 m^3 of rock debris, with an average accretion of 5 cm in the deposit in a long zone along the lower part of the Ridonjira brook (Fig. 5.24). This avalanche was apparently a low-frequency large event, as no significant previous deposit existed and the birch forest was mature in the lower part of the runout zone. In the previous year May 17, a small slush avalanche, reaching down to about 430 m altitude, carried approximately 20 m^3 of rock debris.

A slush avalanche ran on the westfacing slope of Mt. Nissun-tjärro, Nissunvagge, in spring 1982 (for position see Map I). It followed a small brook incised in the mountain wall and continued across a 15° footslope for about 350 m . The debris was thinly deposited over an area of about $30\,000 \text{ m}^2$. Total amount of material is estimated at 250 m^3 .

Other recent slush avalanches of similar geomorphic capacity occurred at Kaisepakke (Pånjenjäske) 1979, Kärkevagge 1980, northern Nissunvagge 1984 and Rakaslako 1985. Avalanches of smaller erosional effect were observed at Kärkevagge 1978, Rakasjokk 1982 and Kårsavagge 1982. Almost clean slab avalanches occurred in 1984 and 1985 at the site near Låktatjåkka (No.19 Fig.5.3) earlier described by Rapp (1960b). The results are summarized in Table 5.12. To be able to check future material transport and deposition, field experiments have been initiated at a number of slush avalanche sites in the Abisko area. Paint markings of stones along the tracks, and partial cleaning and



Fig. 5.24. Dirty snow and slush deposited along the brook Ridonjira, Mt. Njulla, on June 2, 1982. Picture taken three hours after the avalanche occurred. Notice the turbulent stream between the snow banks. Photo O.Nordell.

painting of block surfaces in the runout zones, are used for periodic monitoring of the geomorphological activity of the avalanches.

B. Estimates of postglacial deposition rate

An attempt was made to infer postglacial debris accumulation by slush avalanches at the Kärkerieppe site, central Kärkevagge, by means of an estimate of the thickness of the material. This deposit is superimposed on an alluvial fan with regular morphology, and with the assumption that the fan surface can be extrapolated beneath the avalanche deposit, a gross estimate of the thickness of the latter should be possible. Several pits dug in the avalanche deposit revealed the assumed stratigraphy, with alluvial sediments underlying avalanche transported material (cf. Nyberg 1980). The genesis of the deposit is suggested to be as follows: after a period of combined fluvial/avalanche sedimentation after the deglaciation, strong avalanche activity deposited large amounts of debris causing a shift of the drainage for the

site	No.in Figs. 5.3 & 5.4	time of occurrence							
		1978	1979	1980	1981	1982	1983	1984	1985
Kärkevagge N	12	•		•					
Kärkerieppe	13	•		●					
Kårsavagge	16			•					
Kårsavagge	18					•			
Rakaslako	21			•					•
Låktatjåkka	19							•	•
Rakasjåkk	22					•			
Kåppasjåkk	23					•			
Njulla	24				•	●		•	
Tjuonatjåkka	7			•					
Pånjenjira	14			•					
Kaisepakte (Pånjenjåske)	15		●						
Nissunvagge	2					●			
Nissunvagge	1				•			●	
		● >200 m ³ rock erosion ■ 50-200 m ³ " • <50 m ³ "							

Table. 5.12. Recent observations of slush avalanche magnitude and frequency in the Abisko Mountains

fan sideways. Fluvial sedimentation subsequently continued at lower levels, while avalanches successively increased the height of the part abandoned by the stream channels. Using the mathematical approach of Troeh (1965), the figures for depth of avalanche-transported material above the alluvial fan surface shown in Fig. 5.36 were obtained. They represent the deposition by avalanches after the initial in-filling of a valley bottom of unknown depth by both stream sediments and slush avalanche debris. Assuming intermittent accumulation of avalanche debris during 9000 years in postglacial time, this is equivalent to a rate of accumulation of at least 0.4-0.5 mm/year on the central part of the fan. This value can be compared to reported values for snow avalanche debris accumulation rates of about 0.5-5 mm/year (Luckman 1978).

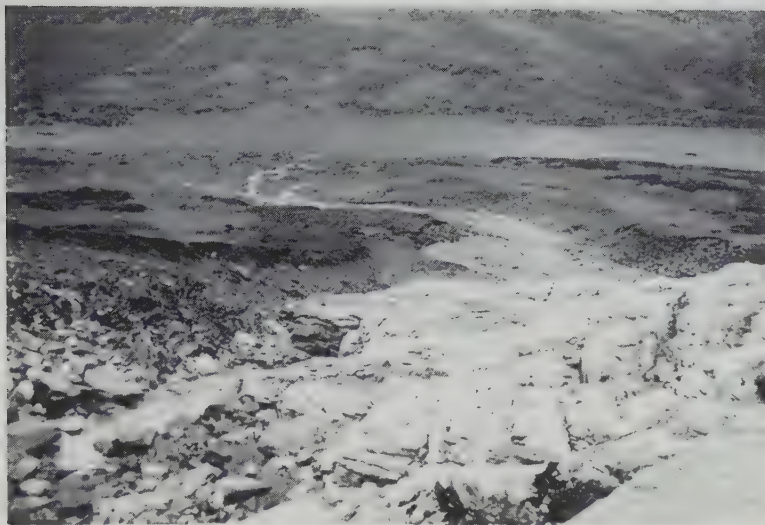
A C14-dating (LU-1847, 730±55 yrs BP, Håkansson 1982) of a buried organic horizon at 0.4 m depth in the slush avalanche deposit by lake Kaskamus Kårsavaggejaure indicates a maximum accumulation rate of 0.5-0.6 mm/year during the last 700-800 years, thus fairly similar in magnitude to the estimate for Kärkevagge. Corresponding rates were obtained with this technique for avalanche cones in USSR, for a period of 1500 years (Miagkov, quoted in Tushinskiy 1966).

5.5.2 Landforms created by slush avalanches

The main geomorphological impact of slush avalanches in northern Lapland is along first-order stream courses, the degree of



a



b

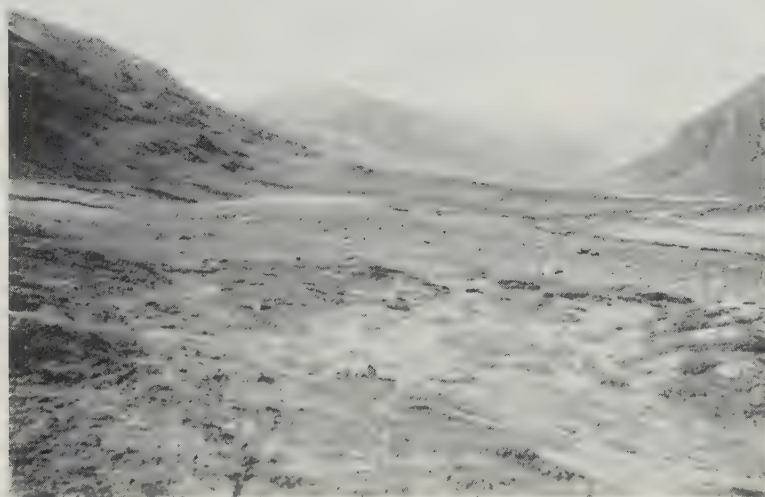


Fig. 5.27. Slush avalanche deposit, Kårsavagge (site no.17, Fig. 5.3.). a. stream diversion caused by the avalanche deposit. Bedrock (dolomite) exposed in the stream channel as a result of avalanche erosion. b. long profile, showing drumlin-like morphology and peripheral block string. Photo 26.8.1980.

Fig. 5.25. Site of removed boulder, about 2x3 m large, in runout zone of slush avalanches, Kärkevagge. The vegetation cover indicates fairly low frequency of avalanches or only localized erosion. Site no. 12, Fig. 5.3. Photo July, 1978.

Fig. 5.26. Debris coating due to slush avalanches along stream course, western slope of Mt. Kaisepakte. The material is dropped by the avalanches after they pass the bedrock threshold in the background. Note the unstable, chaotic nature of the deposit. Photo 30.8.1980.

influence depending on factors such as avalanche magnitude and frequency, debris availability and activity of fluvial processes. As for debris flows, forms resulting from erosion and deposition can be distinguished.

A. Erosional forms

Slush avalanches cause localized erosion along stream channels, resulting in vertical and lateral scouring of the channel. A slush avalanche track along a stream can often be identified on smooth, 'clean-swept' margins bordering the channel and local exposures of the bedrock in a wide zone along the stream course. Shallow grooves a few meters long on vegetation covered or bare slopes are examples of small-scale erosional forms attributed to slush avalanches (cf. Rapp 1960b p. 141). Other features indicating erosion by slush avalanches are surfaces after removed boulders (Fig. 5.25), scoured tree trunks along the channel and impact features formed at a slope foot (see below).

B. Depositional forms

In the runout zone of the avalanches, the debris load from successive avalanches accumulate into debris coatings (Fig. 5.26) or actual depositional landforms. The characteristic unstable positions assumed by avalanche deposited clasts make even very thin or diffuse debris coatings along a stream course diagnostic of slush avalanches. Often a secondary effect of debris accumulation is a stream channel diversion (lateral shifting), cf. Fig. 5.27.

Major depositional landforms created by snow avalanches have been described as avalanche boulder tongues and avalanche cones (Rapp 1959, Luckman 1977, 1978). The former usually occur on talus slopes as a concave basal extension. Avalanche cones occur in front of small stream valleys or gullies in steep mountainous terrain, sometimes below the tree line. They are similar in shape to alluvial cones, but have a different detail morphology, lacking obvious signs of fluvial action (Veyret 1960, Luckman 1978). Impact forms, where the deposit is linked with localized erosion in the runout zone, are special cases that have attracted much attention (Peev 1966, Liestöl 1972, Corner 1980, Hole 1981, Fitzharris & Owens 1984).

Debris deposits created by slush avalanches have been characterized as akin to snow avalanche deposits. They resemble the fan type of boulder tongues (Rapp 1960b) or avalanche cones (reworked alluvial cones, Larsson 1974). A landform probably more unique for slush avalanches is the 'whale-back' fan (slushflow fan) situated in front of stream gullies (Jahn 1967, Raup 1971, Washburn 1979).

Four types of major depositional landforms due to slush avalanches have been distinguished for the study area:

1. Slush avalanche fans: large fan-shaped deposits situated on valley bottoms in front of tracks (small stream valley or gully), with an intervening shallow depression. Their position is often adjacent to or superimposed on an alluvial fan. Typically, the surface is grass-covered with scattered boulders (Fig. 5.27).

2. Multiprocess cones (Slush avalanche/alluvial cones): cone-shaped deposits of combined fluvial and slush avalanche origin, situated at mouths of gullies or stream valleys, usually at the junction between the slope and the valley bottom (Luckman 1978). The surface may be grass-covered but has more or less distinct fluvial channels. A continuum may exist between this type and no.1, depending on avalanche activity (Fig. 5.28).

3. Slush avalanche boulder tongues: flat, tongue-shaped deposits, generally distinctly longer than wide, situated in or beside a stream course used as an avalanche path. The surface lacks or has only scant vegetation, consisting of a mixture of gravel and boulders. Small stream channels may dissect the deposit. Usually not associated with alluvial fans. Seldom as regularly tongue-shaped as boulder tongues beneath talus slopes (Fig. 5.29).

Two varieties occur of type 3 with the following characteristics:

a. stream-crossing tongues: where the path crosses a second stream course in the runout zone, the deposit is situated on the distal side of this usually higher-order stream (Fig. 5.30). Occasionally the morphology of the deposit is ridge-like, with a steep slope facing the stream (cf. Corner 1980).

b. tongues with impact pits: characterized by a water-filled depression created by avalanche impact forces in the proximal part of the tongue (Fig. 5.31).

4. impact pools: where the runout zone ends in a lake, an arched ridge of material is created near the shore by the impact of the avalanches. The ridge encloses a deeper basin or 'pool' (Fig. 5.32).

Terrain profiles of the first three types are shown in Fig. 5.33 and 5.34. Some cases will be described in more detail below.

C. Type cases

Central Kärkevegge, Kärkerieppe cirque

This site, No.13 in Fig. 5.3, is of the 'slush avalanche fan' type, situated on a valley bottom separated from the slope by a small depression (cf. Jahn 1967). Rapp (1985) referred to the deposit as an avalanche boulder tongue, formed by slush avalanches. However, its morphological characteristics, as well as size and terrain position, differ from the typical avalanche boulder tongues that are such prominent features along many high mountain valleys in Lappland. Hence, the term 'slush avalanche fan' or 'slushflow fan' has been preferred here (cf. Washburn 1979).

Some of the characteristic features of the avalanche fan are shown by Figs. 5.35-5.38. The surface is grass-covered apart from scattered blocks with a characteristic concentration in a horseshoe-shaped string (Fig. 5.39). This block line can be used as a rough diagnostic of maximum avalanche reach or runout distance (Rapp 1959 p.44). There are no fluvial channels on the surface. The overall morphology is stream-lined, almost drumlin-like. The distal slope gradient is steeper than on the adjacent alluvial fan and a major difference to this is the

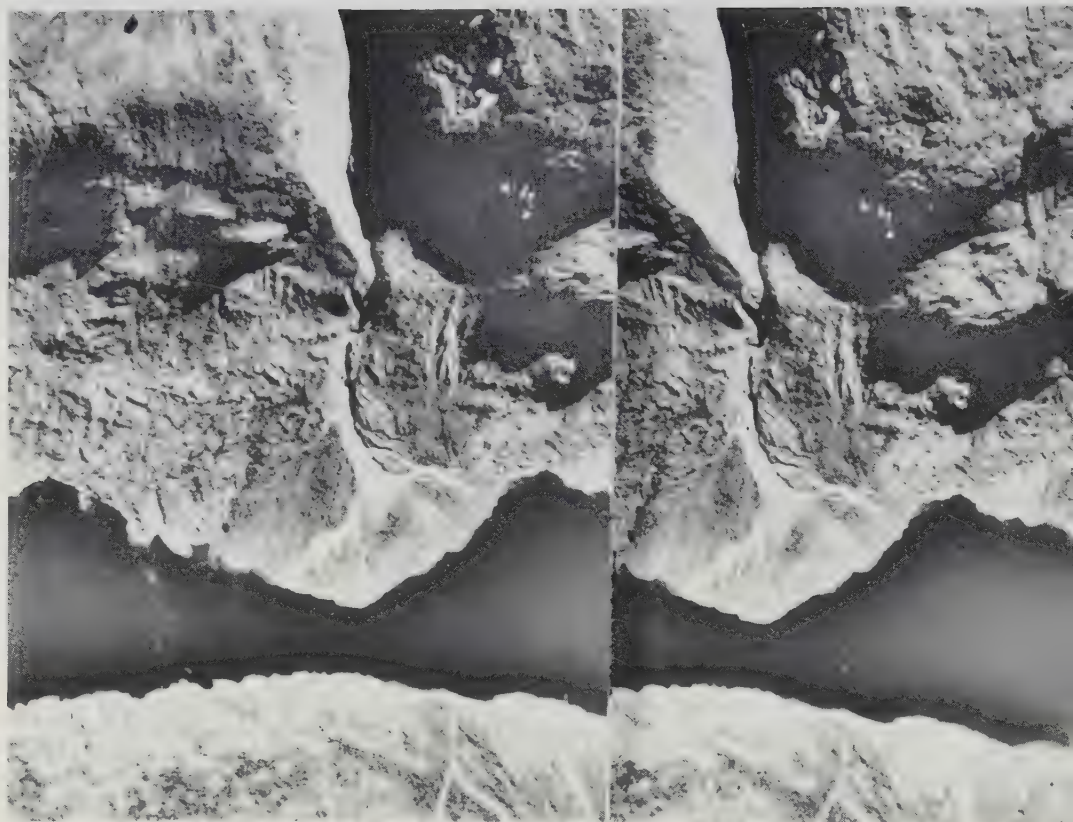


Fig. 5.28. a. Large multiprocess cone on the eastern shore of Lake Njuongerjaure. The cone constitutes an alluvial cone with three parts, one presently inactive and dissected part (left), one part heavily affected by slush avalanches (marked with x), and one part with recent fluvial activity on both sides of the avalanche-affected part. The strong avalanche influence makes the form transitional to a slush avalanche fan. Air photo LMV 1959.

b. Valley-bottom deposits affected by slush avalanches in Sarvesvaggi, Sarek. Three slush avalanche tracks end on the valley floor: one coming from the left, and two from the right (marked by x). The deposits may be classified as multiprocess cone (lefthand case) and boulder tongues crossing a stream (right), in one case associated with an impact pit at the slope foot (arrow). Note the strong influence on the valley bottom drainage lines by the deposition of avalanche-transported material. Air photo A. Rapp 29.7.1967.

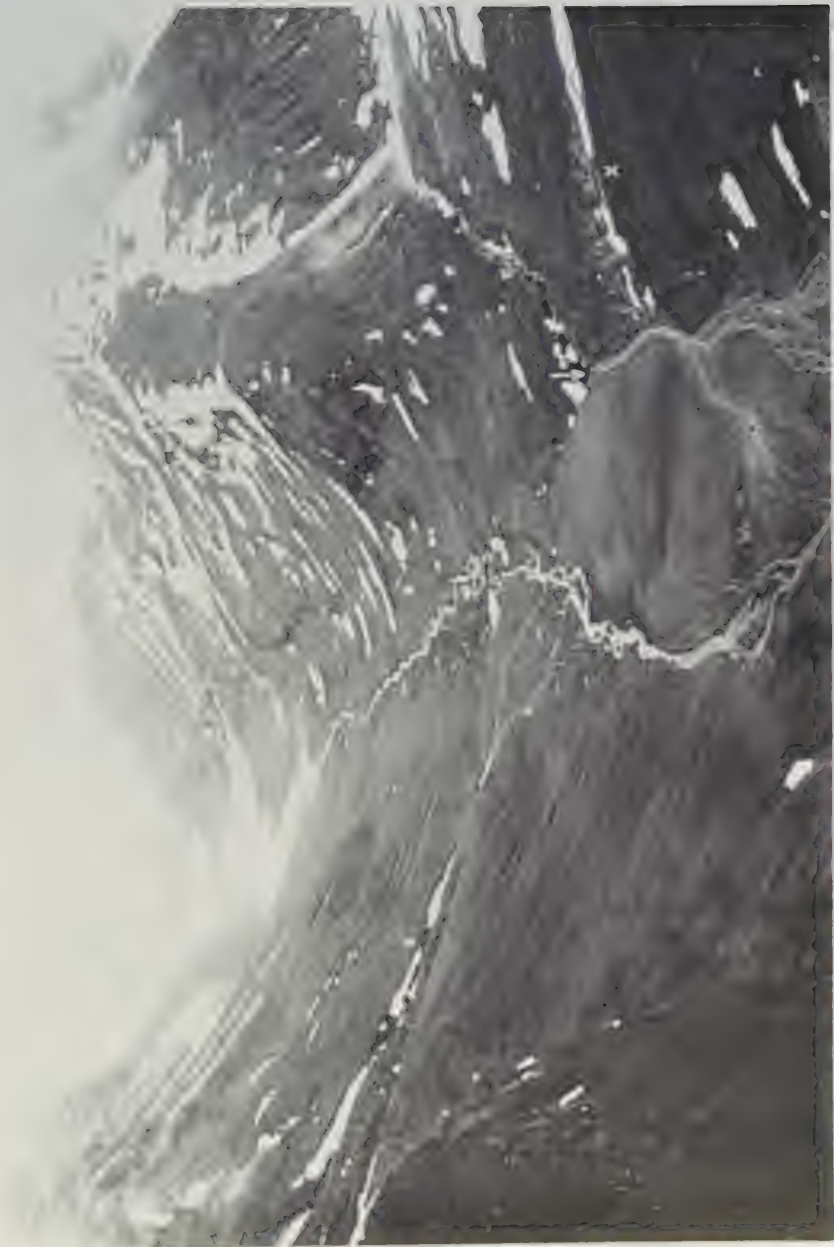




Fig. 5.29. Slush avalanche boulder tongue at Pånjenjäske, Kaisepakte. Dirty snow from an avalanche in May 1979 partly covers the stream bed. Note the irregular detail morphology of the deposit, and its encroachment on the birch forest. Photo 6.6.1979.

irregular morphology which characterizes the highest, proximal part, marked by several transverse ridges, with a local height of about 1 m. Similar ridges occur on some of the other slush avalanche fans in the Abisko area, e.g. in Kårsavagge, where several large fans occur (Fig. 5.40).

Ridges on avalanche deposits have been described for sites with obvious impact effects, where they form distally to an impact pit or stream crossed by the avalanches (Peev 1966, Liestöhl 1972, Corner 1980). For the Kärkevagge and Kårsavagge sites, it seems difficult to explain the ridges by avalanche impact effects. A hypothetical explanation is that the ridges originate as large boulder lines related to avalanches of comparatively short runout distance, subsequently added to and streamlined by later avalanches. High frequency of avalanches would probably obliterate the ridges, so they are suggested to indicate infrequent avalanches with high debris load. The reason why similar features are not known from avalanche boulder tongues would thus be the higher frequency of snow avalanches compared with slush avalanches, with a usually much lower debris load of individual avalanches, preventing the accumulation of suitable nuclei for ridge development.



Fig. 5.30. Slush avalanche boulder tongue deposited on a reverse slope with respect to the avalanche track, which crosses the valley bottom stream. Livamvagge west of Lake Livamjaure. Numerous snow avalanche paths visible on the mountain wall south of this lake, to the left in the background. Photo 17.7.1977.

Kaskarieppe, Nissunvagge.

An example of a slush avalanche-affected alluvial cone or a multiprocess cone can be found by the lake 1140 in the westerly tributary valley to Nissunvagge. The site is No. 4 on the map, Fig. 5.4. A detail map of this site has been made photogrammetrically (Fig. 5.41). The locality shows the marked difference between slush avalanche and snow avalanche tracks. The rock wall west of the valley has a series of well developed avalanche boulder tongues at its foot, mainly of the roadbank type. The slush avalanche track is along a small stream draining a glacier and flowing into the lake from the south.

The multiprocess cone rises 20 m from the lake. The largest slush avalanches reach the lake shore, as revealed by a distinct block-string along the cone periphery near the shore. In contrast to the Kärkevagge slush avalanche fan, shallow fluvial rills and channels occur on the cone surface, showing that fluvial activity is important and in this case dominant over the avalanches in the development of the cone. There is no depression at the apex of the cone separating it from the slope, as for the slush avalanche



Fig. 5.31. Avalanche impact pit, west of Mt. Atjektjåkka (north of St. Sjöfallet area). The fairly gentle slope gradients imply that slush avalanches are the active process forming the pit. Air photo LMV 1959.

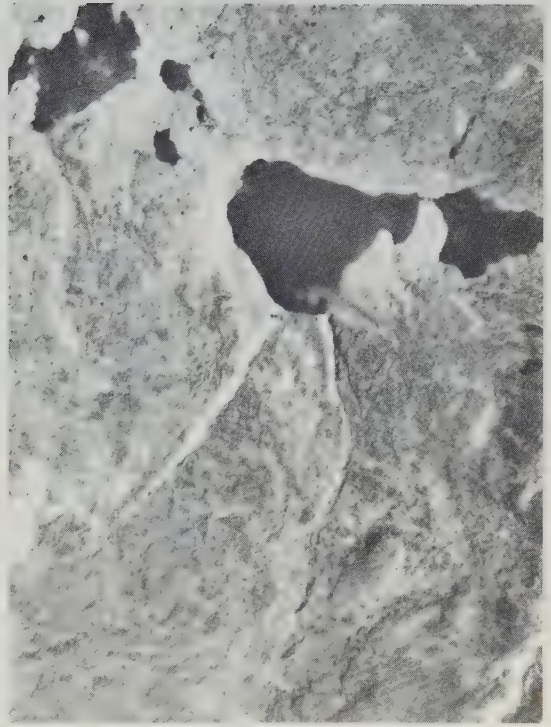
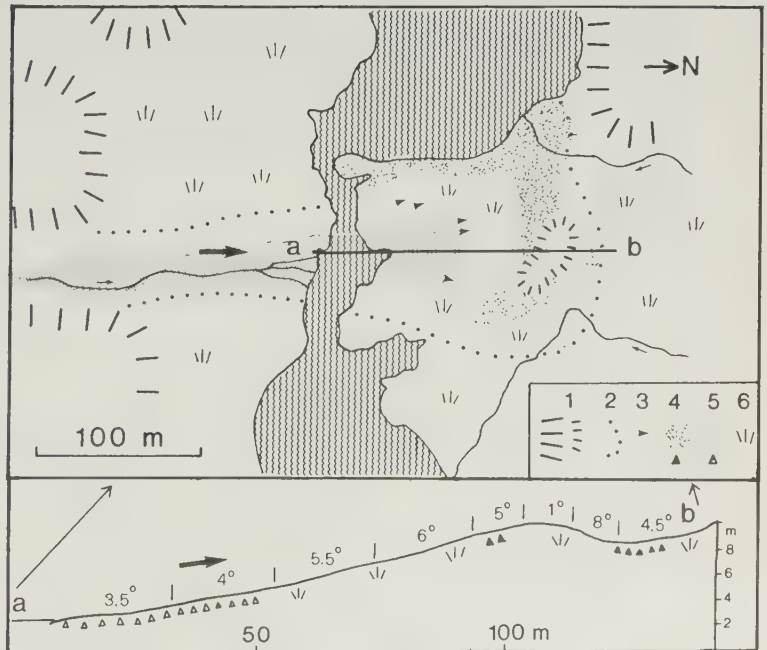


Fig. 5.32. Two avalanche impact pools at Lake Råggejaure (the arched ridges on the lake bottom in front of the two stream courses). As the surrounding slopes are of fairly gentle inclinations, it is likely that the pools are due to the impact of slush avalanches, rather than snow avalanches. Air photo LMV 1959.

Fig. 5.33. Slush avalanche boulder tongue deposited across valley bottom stream in Livamvagge. Key:

1. slope of hummock
2. runout zone of avalanches
3. avalanche debris tail
4. block field with old lichen cover
5. do with young or no lichen cover
6. grass cover

Note the capacity of the avalanches to transport blocks more than 100 m upslope.



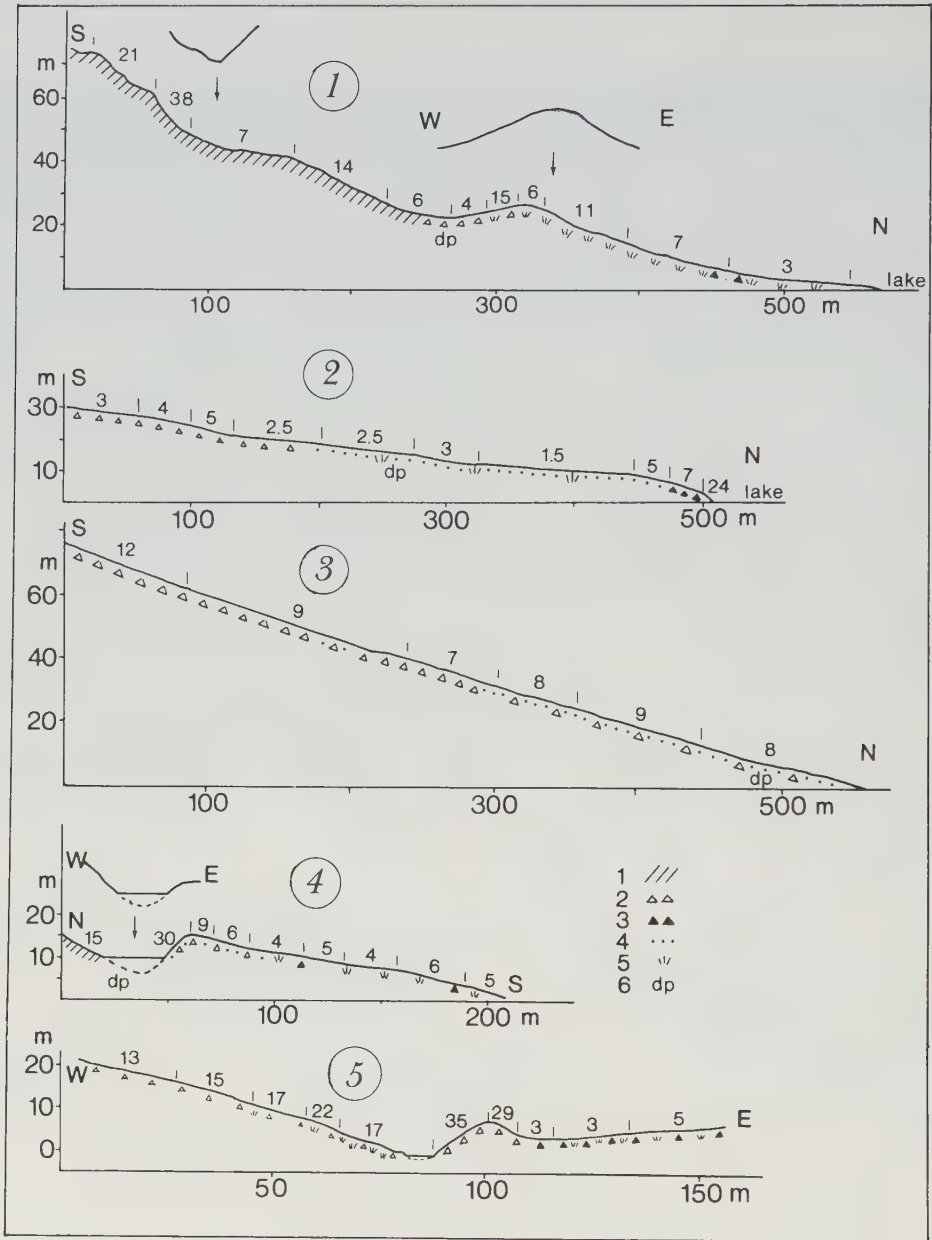


Fig. 5.34. Terrain profiles of slush avalanche paths or runout zones. 1. slush avalanche fan, Kårsavagge (No.17, Fig.5.3). 2. multiprocess cone, Kaskarieppe, Nissunvagge (No.4, Fig.5.4). 3. slush avalanche boulder tongue, Pånjenjåske, Kaisepakte). 4. avalanche impact pit, Kuotekjaure. 5. slush avalanche boulder tongue across stream, Leirdalen, Jotunheimen, Norway. Key: 1. bedrock outcrop 2. lichen-free boulders 3. lichen-covered boulders 4. sandy-gravelly deposit 5. grass cover 6. diversion point of stream

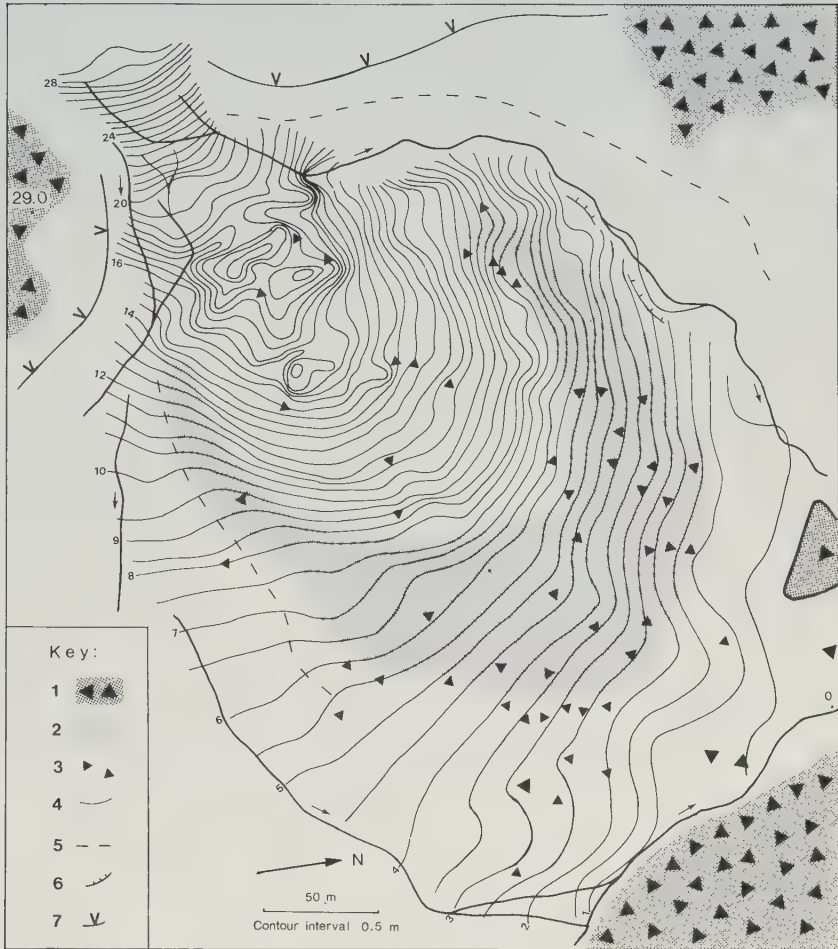
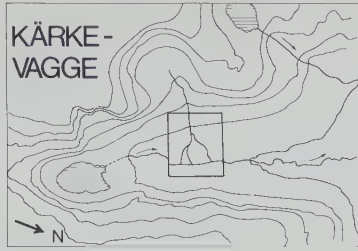


Fig. 5.35. Contour map of the slush avalanche fan, central Kärkevagge. Key: 1. deposit of large boulders of probable late-glacial age 2. area of block concentration by slush avalanches ('block string') 3. boulder larger than one metre 4. active stream 5. inactive stream 6. fluvial erosion scarp 7. foot of major slope. Relative heights in m.

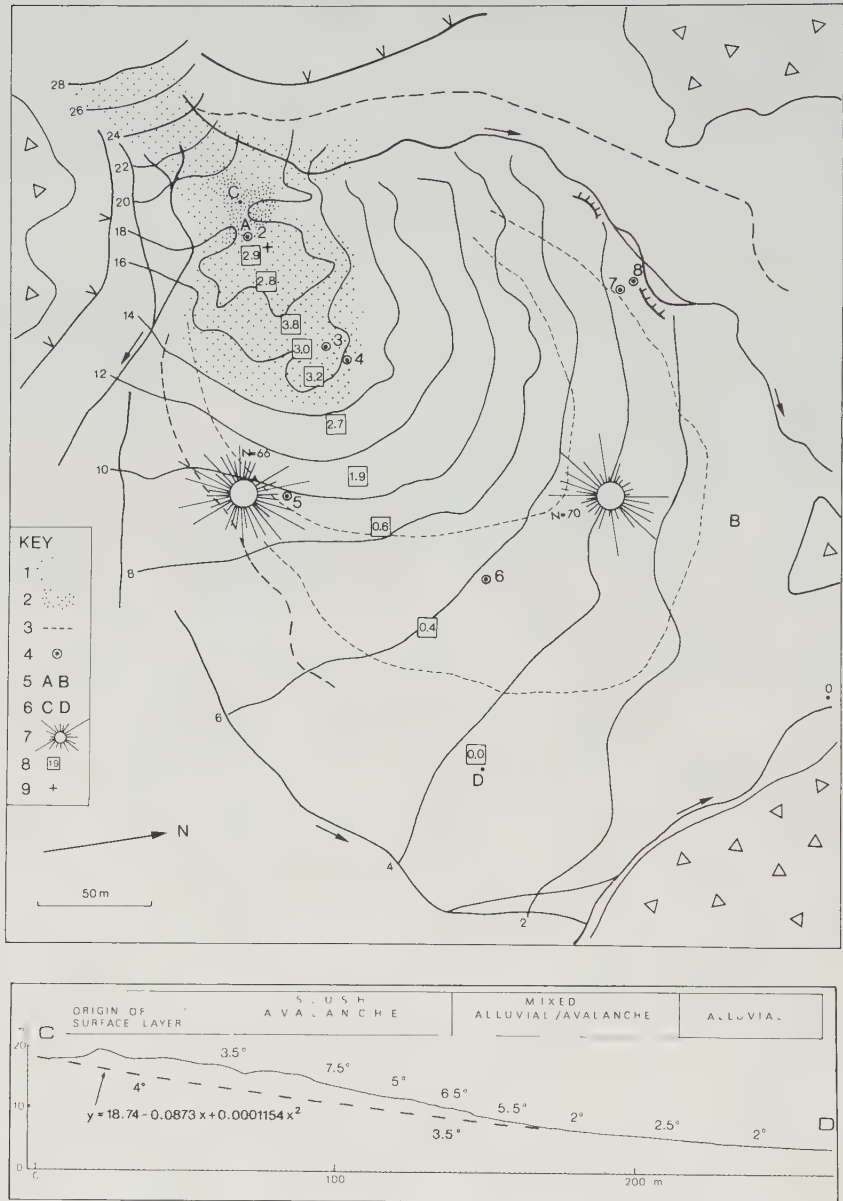


Fig. 5.36. Recent additions of material in proximal part of the slush avalanche fan in central Kärkevagge, from 1980 (key:1), and 1978 (key:2). Position of block string marked by finer stippled lines (key:3). Orientation diagrams (7) for surficial boulders show large spread in directions, with a tendency to long-axis alignment with avalanche movement especially in distal central part of the string. Numbers in squares (8) denote estimated minimum depth of avalanche material in m (cf. text). Other features according to key: 4. soil sample site 5. profile for lichen measurements (cf. Fig. 5.47) 6. profile for mathematical derivation of depths of avalanche material 9. boulder used as fix point in plane table mapping. For remaining symbols, see Fig. 5.35.

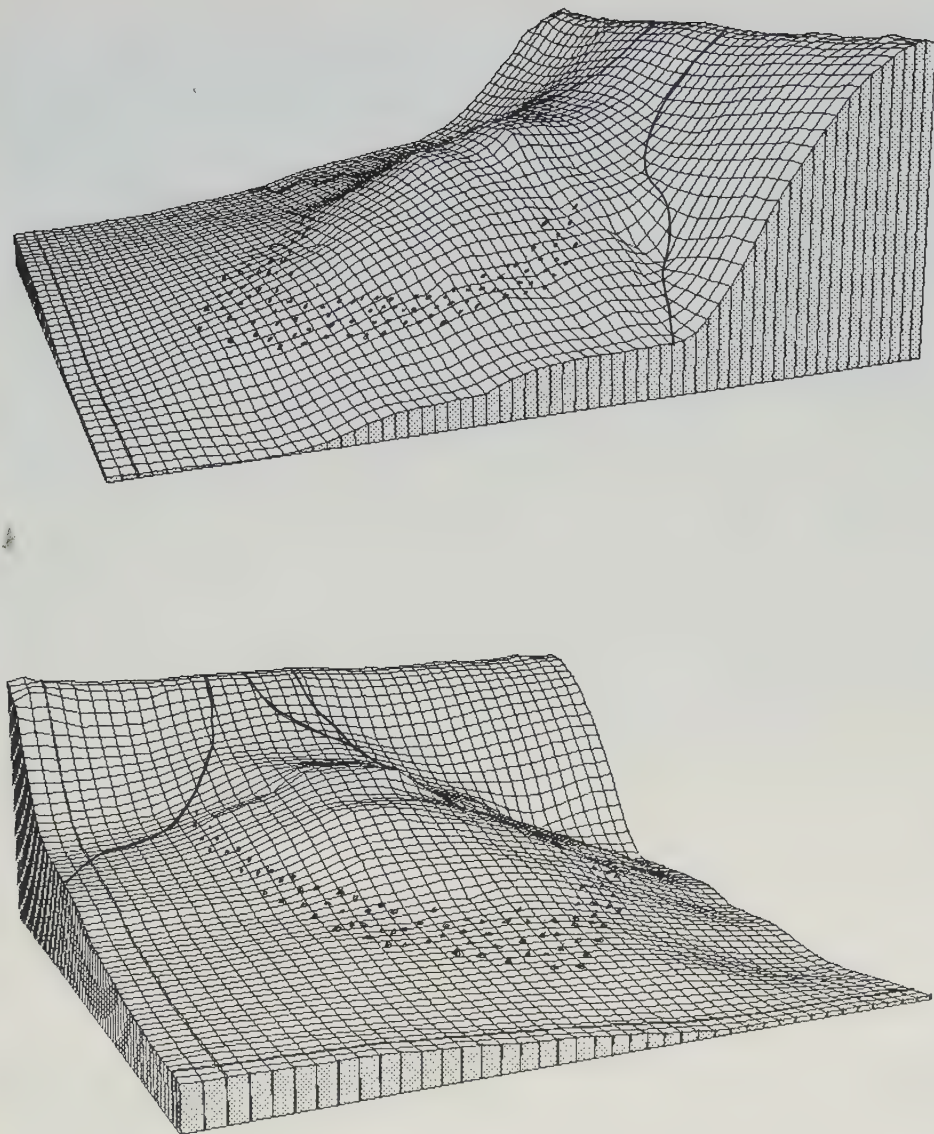


Fig. 5.37. Digital terrain models of the slush avalanche fan, central Kärkevagge, seen from the north (above) and east. Position of block string indicated by points, streams by thick lines. Note transverse ridges and proximal depression, which are features that distinguish this type of debris accumulation from alluvial fans. Vertical exaggeration about 2 x. For scale of the fan, see Fig. 5.36.



Fig. 5.38. The slush avalanche fan of central Kärkevagge, seen from Mt. Vassitjåkka. The horse-shoe shaped block string marking the range of the largest avalanches is clearly seen, pointing towards the research hut of the valley. Photo August 1979.



fan, so the shifting of stream channels over the whole fan area can still operate. An example of a transitional form between this type and a slush avalanche fan is shown in Fig 5.28.

Pånjenjáske, Kaisepakte.

This deposit, classified as a slush avalanche boulder tongue, is situated about 20 km south-east of Abisko near the Kaisepakte railway stop along Lake Torneträsk (cf. map Fig. 3.1). The avalanches originate in the rock canyon Pånjenjáske of Mount Kaisepakte, a plateau-shaped granite massif rising 865 m above the level of Lake Torneträsk. The latest avalanche occurred in 1979 (Fig. 5.29).

Below the canyon, the path follows a channel incised 2-4 m in till for more than one km downstream, where it ends in a deposit of bouldery gravel encroaching upon the birch forest at 530 m altitude (Fig. 5.42). The terrain position is at a stream bend, a typical location for many slush avalanche boulder tongues. The slope gradient at this point is about 8 degrees. Immediately upstream of the deposit, the track crosses the foot-path from Kaisepakte towards Lake Pessisjaure.

The boulder tongue at Pånjenjáske is very different from the two cases described previously. It has a more irregular surface shape and lacks vegetation cover almost completely. The general outline is tongue-shaped, with protruding lobes in terrain depressions. The avalanche origin of the deposit is evident from the fine material present on top of large boulders, where it could not have been brought by stream action.

The chaotic nature of the deposit clearly distinguishes it from debris flow deposits (cf. ch.4). On air photos, this difference may be difficult to observe, and this slush avalanche path, like some others in northern Lappland, has been mapped as debris flow on the geomorphological map (Melander 1977a).

The stream channel downslope of the canyon dissects a large fossil alluvial fan. Parts of this fan were previously affected by slush avalanches. Evidence for this is provided by an avalanche debris tail in the birch forest, formed in lee position to the avalanches behind a boulder. The debris tail is grass-covered, lacks signs of recent avalanche influence and occurs among mountain birches of a height of a few metres.

Kuotektjåkkå

This site provides an example of a slush avalanche deposit with an impact pit. It occurs on the southwestern slope of Mount Kuotektjåkkå (1514 m), a granite massif north of Lake Sitojaure, close to the eastern boundary of the Sarek National Park. Along

Fig. 5.39. Block string on the slush avalanche fan, central Kärkevagge. The avalanche track is to the right of the picture. View towards the south. Scale of avalanche-transported blocks given by reindeers in the background. Photo 3.7.1977.

the chute-sculptured mountain wall towards Lake Kuotekjaure, snow avalanches operate together with rockfalls and debris flows in the formation of debris cones of coarse material (Fig. 5.43). These deposits may be contrasted with the two debris accumulations north of the lake, which reach further out on the valley floor and are associated with large scars or gullies, less steep than the chutes in the rockwall, in the slope to the north-east (Fig. 5.44).

The northerly stream gully originates at about 1300m altitude in two branches. The longitudinal profile is stepped, with two small waterfalls, and a steep lower section ending abruptly at the slope foot, where a large impact pit is situated and distally of this, a debris tongue. Below the upper waterfall is a miniature-form of impact pit, a few meters in diameter. The general appearance of the valley bottom deposit and the main erosional pit is shown in Fig. 5.45. The impact pit is a water-filled depression about 30 m long and approximately 4 m deep (Fig. 5.34). Its distal slope rises 4 m from the water surface and is covered with stones and blocks of fresh appearance, lacking lichen cover. Avalanches periodically scour the depression, since it would otherwise, with its position as a



5.40 b

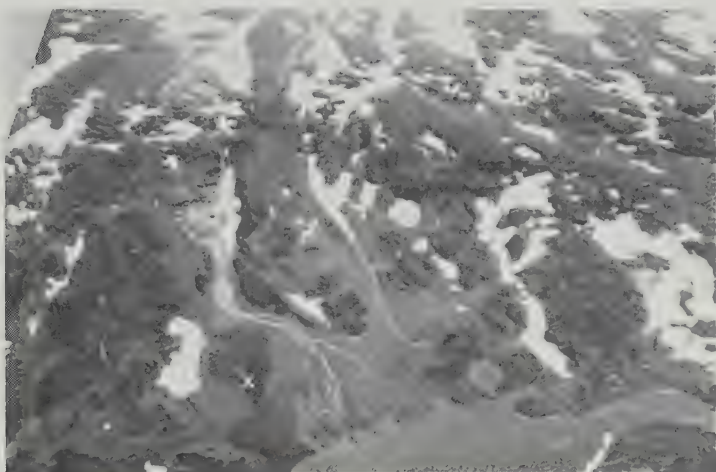
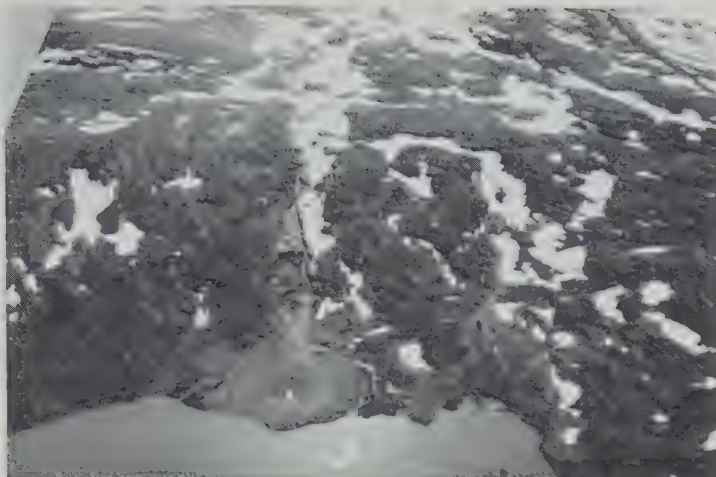


Fig. 5.40. a. Two slush avalanche fans (marked by x) in Kårsavagge (no. 18 and 17 in map, Fig. 5.3). Note the exposed bedrock along the tracks and the diversion of the streams caused by the accumulation of avalanche debris. Transverse ridges are faintly visible on the upper photo, but lacking in the other case shown. Extreme range of the avalanches indicated in upper photo by subaquatic block string, distinguished by reflections on the lake surface. Air photo 12.7.1982.

b. Case no. 18 seen from the southern valley side. Notice the irregular morphology with ridge features, and subaquatic distal block string in the lake. Photo August 1979.

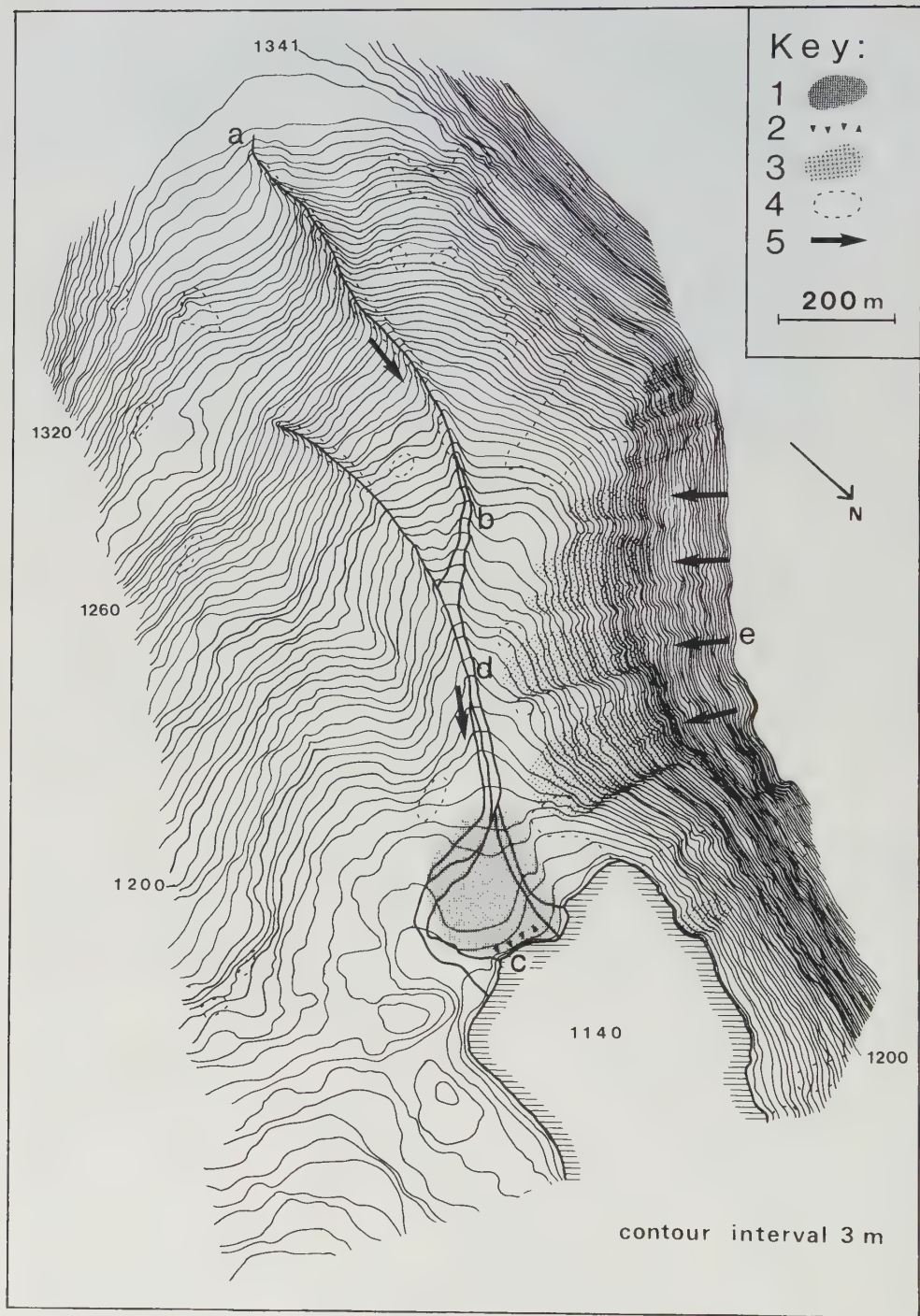


Fig. 5.41. The slush avalanche site at Kaskarieppe, Nissunvagge (a - c) and lower parts of snow avalanche paths on Mt. Pallentjåkka (e.g. d - e). Key: 1. multiprocess cone (alluvial - slush avalanche) 2. block string 3. avalanche boulder tongue due to snow avalanches 4. late-lying snow field 5. avalanche tracks. The map shows the marked difference in terrain localization and slope gradient for slush and snow avalanche paths. Cf. cover photo, taken at about 1270 m altitude along the slush avalanche track, looking northwards.



Fig. 5.42. The slush avalanche track downslope of the Pånjenjäske canyon, Mt. Kaisepakte. The terminal deposit is seen beyond the stream bend, marked by x, about 1.2 km distant from the canyon and photo site. The stream has cut its way through the dirty snow deposits of the avalanche in 1979. In the background, Lake Torneträsk, with rotten ice. Photo 6.6.1979.

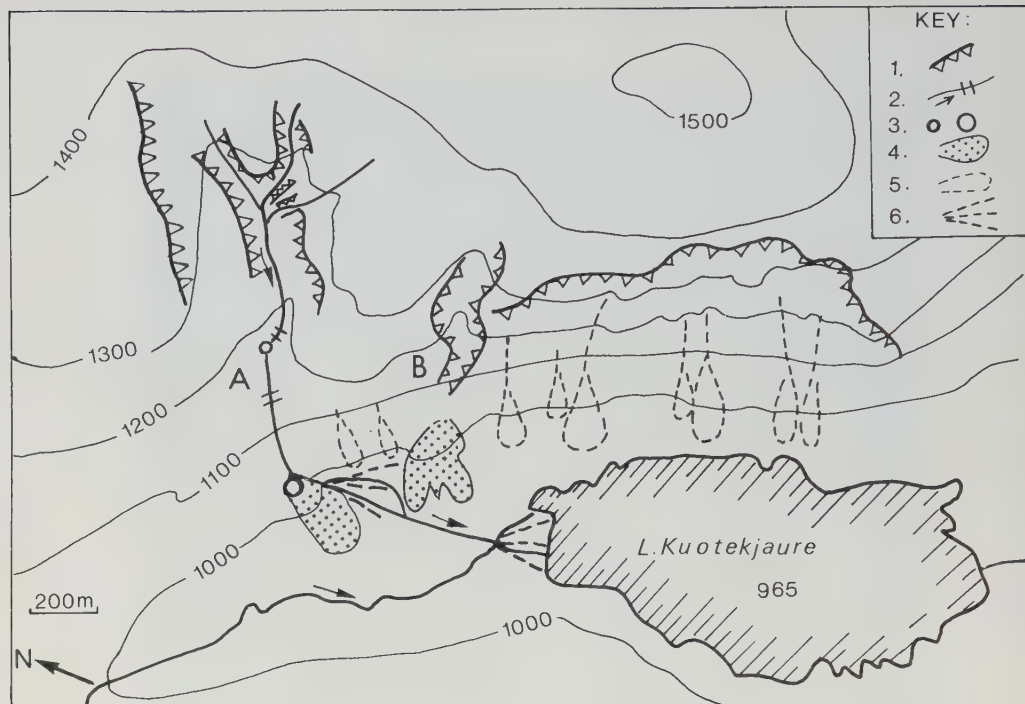


Fig. 5.43. The Lake Kuotekjaure area. The westfacing slope east of the lake is characterized by basal debris accumulations due to rock-falls, snow avalanches and debris flows. Further to the north, large bedrock gullies are associated with debris accumulations situated in the valley bottom. Slush avalanches are thought to be the main active process in the formation of these accumulations. Cf. text. Key: 1. Steep bedrock slope 2. stream with water-fall 3. avalanche impact pits, small and large 4. debris accumulation due to slush avalanches 5. ditto, due to other mass movement processes 6. alluvial fan.

sediment trap along the stream course, soon be eliminated by fluvial sedimentation during rainstorms. Some markings made with the aim of estimating present-day avalanche activity at this site will help verify this assumption later on.

Other cases of impact pits found in the study area have more gentle path gradients, which leave no doubts about their slush avalanche origin.

Räggejaure

As shown by Corner (1980), it is not unusual in steep Norwegian fiord terrain to find avalanche paths ending below the water-line in entrenched pools, scoured by the avalanches. Perhaps many of these steep avalanche paths are due to snow avalanches, although some follow mountain brooks and may be of the slush variety. Two avalanche impact pools are situated by Lake Räggejaure in the southern part of the Norrbotten mountains. The avalanche tracks, about 300 m long, lead down to the lake on the southwestern side (Fig. 5.32). The starting zones are shallow depressions, which make up the headwaters for the small streams followed by the



Fig. 5.44. Rockwalls and debris accumulations north and east of Lake Kuotekjaure. The accumulations in the foreground and in the middle of the picture are due to slush avalanches, the debris slope in the background mainly to rockfalls, snow avalanches and debris flows. Bedrock granite. Photo 3.9.1981.



Fig. 5.45. Avalanche impact pit (water-filled) and debris tongue N of Lake Kuotekjaure. The avalanches follow a small stream gully, the lower part of which is seen in the left part of the picture. The stream is diverted at right angle to its previous course at the impact pit. Size of pit 30 m across. Cf. Fig. 5.34. Photo 3.9.1981.

avalanches. Slope gradients are too low for snow avalanches (about 15°) so with fairly large confidence it can be assumed that the tracks are due to slush avalanches. Two arcuate ridges have been formed subaquatically by avalanche impact, faintly visible in the photo of Fig. 5.32.

D. Influence on slope morphology

In chapter 4, the influence of debris flows and snow avalanches on talus slopes was discussed. Similar effects on talus slopes by slush avalanches have been reported by Caine (1969) and Bones (1973), although possibly they referred to wet snow avalanches rather than slush avalanches of the type dealt with here. In Nissunvagge, the main impact on slope morphology by slush avalanches is along stream channels. In the long run, avalanches may cause modification of channel shapes in arctic streams. An increased rate of channel incision along the avalanche tracks will be combined with deposition along gentle reaches or on footslopes, leading to slope decline.

5.5.3 Summary

The type cases treated and the survey of sites of geomorphic importance in N. Lapland (section 5.4.) both testify to the considerable geomorphological efficacy of slush avalanches. At present however, the number of observed avalanches is too low to allow an estimation of the average amount of erosion with any confidence. The comparative lack of observations of recent avalanches with no erosive effects (excepting cases on glaciers) strengthens the often quoted view that slush avalanches generally have high geomorphic power. This is particularly true of their transporting capacity (Fig. 5.25).

However, in comparison with the large debris flows described for the Nissunvagge area, slush avalanches have moderate total erosional effects. The landforms resulting can be distinguished from the corresponding forms due to snow avalanches mainly by the different terrain positions, associated with streams. The large grass-covered slush avalanche fans described are probably unique forms related only to this type of avalanche. Despite generally low path gradients, also slush avalanches are capable of producing impact forms.

The geomorphological action of slush avalanches along stream courses can be summarized as resulting in channel modification by debris removal and bedrock exposure, and debris deposition, occasionally coupled with impact erosion or channel diversion in the lower reach. The overall effect is similar to that by large, valley-confined debris flows (e.g. Okuda et al. 1980), but with different detail morphology of resulting landforms. Also like debris flows below major chutes, slush avalanches recur on the same site repeatedly, building deposits of large size. Their association with streams may lead to a certain direct link to catchment denudation, but frequently the transported debris is stored out of reach for stream action, and constitutes local redistribution.

5.6 Frequency of slush avalanches in northern Swedish Lapland

5.6.1 Previous information on slush avalanche frequency

Little is known of the frequency or recurrence interval of slush avalanches at particular sites. Some historical data are available, where the avalanches have caused accidents or destruction of human property, e.g. from Iceland (Jónsson 1957), Spitsbergen (Jahn 1967) or northern Scandinavia (Rapp 1960b). For the latter area, the records stem from the railway authorities, who have data stored regarding traffic disturbances along the line Kiruna-Narvik. In these records the cases associated with slush avalanches can generally be separated from other types of mass movements, especially snow avalanches, on the basis of time of occurrence. Slush avalanches usually occur from the middle of May to the beginning of June. Snow avalanches are most important during the period March-April (Schmacke 1959).

The railway records indicate a low frequency, with recurrence intervals of sometimes several decades, e.g. at Mt. Kaisepakte east of the Pänjenjäske track described earlier (cf. Rapp 1960b). However, in the Stora Sjöfallet area, where avalanches cause road disturbances, some slush avalanche tracks appear to be activated more often, but not yearly (Larsson 1974). Unfortunately, apart from the isolated and areally limited study of Larsson, the observations on slush avalanche frequency are not systematic and only refer to cases of practical importance. Thus, they may have given a somewhat false impression of the actual frequency, failing to take also smaller avalanches into account.

5.6.2 Recently observed slush avalanches in the Abisko area

During the years of the present study, from 1977 to 1983, a number of slush avalanche sites in the Abisko region were systematically observed each year. This was done either in the spring or early summer, when disturbance of the snow cover could be checked from a distance or aircraft (in 1983), or later in the summer by inspection at each site for fresh deposits. The result of this study was presented in Table 5.12. The implication of the data is that the frequency of slush avalanches in this area may be fairly high at certain sites.

5.6.3 Assessment of slush avalanche frequency from stratigraphic and lichenometric data

Several of the slush avalanche deposits in the study area are covered with vegetation at the present, usually grasses and herbs but occasionally willow (*Salix* sp.). This indicates a low avalanche frequency at these sites, or mainly non-erosive avalanches. The possibility of finding old buried organic horizons in deposits of this type was investigated by digging of sections. The locations studied were at Kårsavagge (17 and 18), Kärkevagge (13) and Tjuonatjåkka (7), cf. Figs. 5.3 and 5.4. The stratigraphy of these sections was similar in all cases, consisting of an unsorted mixture of angular rock debris with high percentage of stones. The type of material is illustrated in Fig. 5.46. No stratification was evident, and an organic horizon was only found in one case (No.17), at 40 cm depth. The C14-age was 730±50 years BP. Below the horizon, the material was again typical avalanche deposit.

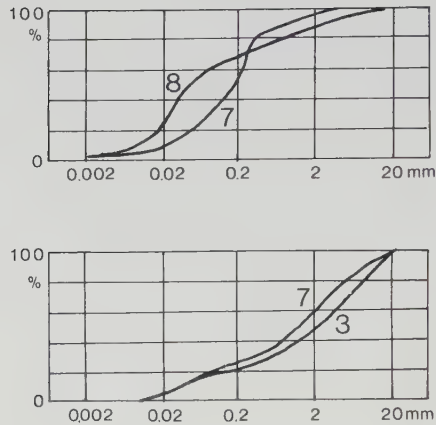


Fig. 5.46. Grain size distribution of samples of alluvial and slush avalanche debris, Kärkevagge fan. Cf. Fig. 5.36. for location. 3=central part of deposit, 0.5 m depth. 7=peripheral part, 0.7 m depth (alluvial material, upper diagram), 0.4 m depth (avalanche material, lower diagram). 8=peripheral part 0.2 m (alluvial material). Notice the poor sorting of avalanche-transported material compared with fluvial sediments.

The horizon may be interpreted as representing a period of low avalanche activity and favourable conditions to plant growth prior to the avalanche event that buried it. A period of warm climate indicated by high tree limit probably occurred about 1000-800 BP (Karlén 1976). However, the dating can only be used for a gross estimate of total subsequent debris accumulation by avalanches (p. 147), and gives no information on the frequency of avalanches during periods when no substantial vegetation cover existed at the site. Furthermore, the approximate nature of any estimate based on only one dating hardly needs to be pointed out.

The lichen cover of blocks deposited by slush avalanches is often well developed on outer (distal) parts of the accumulations. Closer to the avalanche track, the blocks have small lichens or lack them altogether, giving a fresh impression. This indicates a low frequency of large avalanches at present. The variation in lichen size (Rhizocarpon section Rhizocarpon and section Alpicola) was measured on all blocks along a profile of the slush avalanche fan in Kärkevagge (before the latest avalanche occurred in 1980). Fig. 5.47 shows the clear increase in lichen size in the outer part of the fan. Blocks of the central part had lichen sizes indicating tentative maximum ages of 50-60 years, whereas blocks of the outer part were up to 300 years old, according to the lichen sizes (cf. Karlén 1973). The difference cannot be explained by snowcover effects or other environmental influences, but must reflect the time elapsed since the blocks were deposited by avalanches. Of course, isolated recent avalanches may be responsible for the small lichen sizes measured in the outer part, but it seems reasonable to conclude that the frequency of major avalanches has decreased since the latter part of the 17th century or that they have recurrence intervals measuring several hundred years.

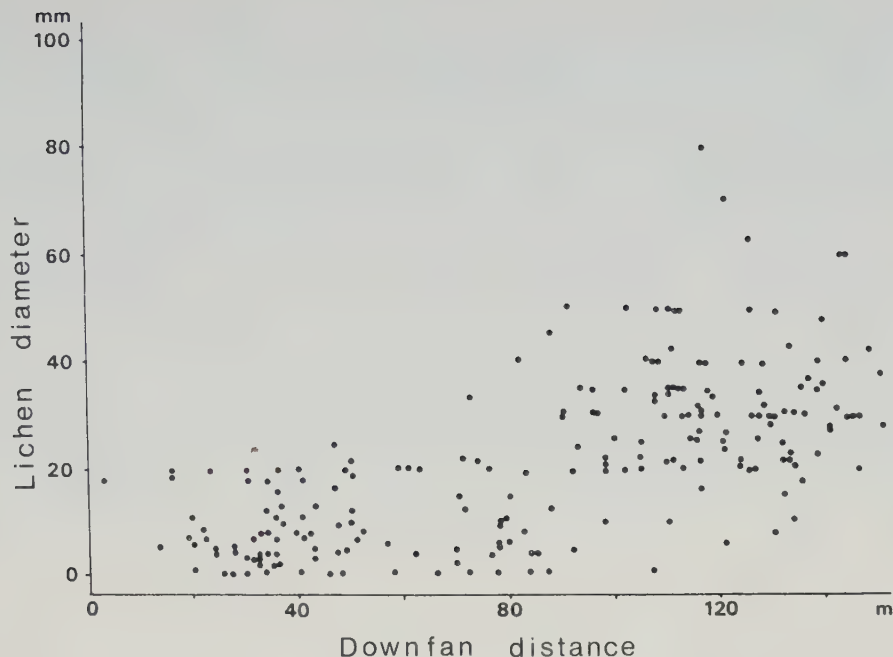


Fig. 5.47. Size frequency distribution of lichens (Rhizocarpon section Rhizocarpon and section Alpicola) on blocks along a profile on the slush avalanche fan, central Kärkevagge. For location of the profile, see Fig. 5.36. The increase in lichen size downfan is interpreted as a result of either a very long recurrence interval of large avalanches reaching the peripheral part of the fan, or a lowered frequency of large avalanches during the last 200-300 years.

5.6.4 Other methods of assessing slush avalanche frequency

Where slush avalanches affect birch forest, dendrochronology may provide some information on frequency. The avalanche at Njulla in 1982 felled mature birches along its path (Fig. 5.10). Tree-ring datings of a few of the largest trees damaged by the avalanche give ages of at least 50-60 years, indicating no avalanche of similar magnitude in this time period. A Salix bush growing on site 18 Kårsavagge and sampled in 1979 was dated by dendrochronology to about 15 years of age. However, it is doubtful whether this can be taken as evidence for no avalanche occurrence in this period, as observations of the recent case (1982) at the site, showed that low bushes may survive overriding by an avalanche.

Climatic records might be used to estimate frequency of slush avalanches, if more reliable data were available on the significance of various factors, e.g. temperature, insolation and snow cover depths, in causing release of the avalanches. However, the problem mentioned in section 4.4 concerning debris flows, namely the uncertainty in extrapolating meteorological station data from low altitudes to the high mountain areas, is valid also for slush avalanches.

5.6.5 Summary

The data presented concerning temporal occurrence (frequency) of slush avalanches in northern Lappland may be summarized in the following way:

1. Observations in the Abisko area indicate a large variability in frequency for different sites during the period 1977-1984. Some sites have had no avalanches, whereas other experienced 2-3 cases in this short period of time. The timing of the avalanches seems to be partly influenced by local site factors and only in some years (e.g. 1982) due to some common external cause of meteorological nature.
2. The present vegetation, especially lichen cover, on slush avalanche deposits indicates a lower frequency of very large avalanches during perhaps the last 300 years than earlier. Possibly a low frequency also characterized a period prior to about 700-800 years BP, allowing the establishment of a vegetation cover at avalanche sites.

5.7 Regional distribution of slush avalanches

5.7.1 Available data

As slush avalanche tracks and deposits are not included on the geomorphological maps of the Swedish mountain area, the information available prior to this study covered only a small number of cases (Rapp 1960b, Larsson 1974, Nyberg 1980). The data collected in the present study are based on air photo interpretation, and consequently embrace only slush avalanches with marked geomorphological impact. As for debris flows, there is a possibility that very old forms, presently inactive, may be poorly represented, although the generally larger size of slush avalanche landforms compared with debris flows probably makes this a less serious deficit in the total picture of the distribution of slush avalanche sites. Neither are there any gaps in the distributional pattern due to overlapping with other forms, as for debris flows on the geomorphological maps. The accuracy of the air photo survey was checked during field traverses (Fig. 3.1, more than 300 km in total length). 27 of the sites mapped were visited and found to be correctly interpreted. 3 sites missing on the maps were encountered. Hence, extensive field inventory would probably add some further sites, but this would not radically alter the result of the survey.

5.7.2 Distributional patterns

The result of the air photo survey is shown in Fig. 5.48 and in overview on Map III. More than 200 sites were found, giving an average density of 0.75/100 km². Although the majority of slush avalanche sites were found in the high relief areas, there is no marked association between the distribution and these areas, as for debris flows and snow avalanches. In part this may reflect the low slope angle required to release slush avalanches. Very steep terrain favours frequent snow avalanching and therefore has less potential for the building-up of deep snowpacks linger-

ring into the spring melt season. A lack of topographic thresholds in steep terrain may also contribute, by not offering suitable starting zones, to the relatively seen moderate concentration of slush avalanches to high relief areas. In areas dominated by gentle slopes, slush avalanches may be the most important type of rapid mass movement, other erosional processes of similar magnitude being absent because they require steeper gradients to be initiated.

Compared with debris flows, slush avalanches have a much more limited spatial geomorphic impact, strongly linked with the existing drainage network. The clustering of sites is not as marked as for debris flows, and there need not be any association between adjacent sites regarding time of initiation or repeated occurrence. Thus, the distributional pattern appears to be largely a result of an interaction of local terrain conditions conducive to slush avalanching, and the incidence of snow melt events causing activation for particular sites. As the recurrence interval of such events seems to be fairly short (cf. section 5.6), it seems likely that most potential sites have been active several times in postglacial time.

5.8 Slush avalanches as natural hazards

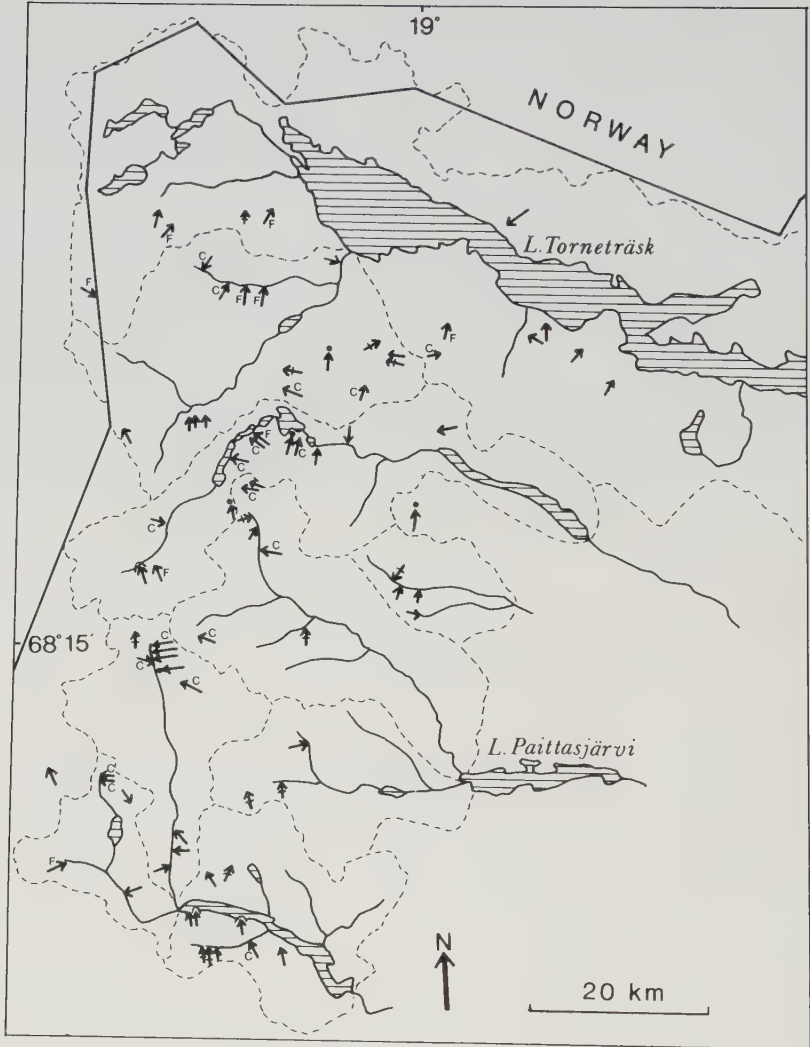
5.8.1 Hazardous terrain sites for slush avalanches

Unlike debris flows, slush avalanches in Swedish Lapland are strongly localized processes linked with stream channels, occurring elsewhere probably only on glaciers. The danger of snow avalanches on steep mountain slopes is fairly well known, but knowledge of the risk for slush avalanches along stream courses is almost non-existent. Many people have been killed by slush avalanches in Iceland, to judge from descriptions in Jónsson (1957), cf. also Björnsson (1980). Very wet avalanches have been reported to cause damage to human life and property also in Norway (Domaas & Lied 1981). In Sweden, two areas have experienced problems with slush avalanches causing destruction or blockage of roads or foot paths and presenting an obvious hazard to people: the St. Sjöfallet area (Larsson 1974) and Mt. Njulla near Abisko (Nyberg 1982).

From the survey of slush avalanche sites made in the present study, it is evident that no typical features exist that particularly distinguishes certain stream courses from others with respect to slush avalanche hazard. Possibly the presence of a steep waterfall or threshold in a stream course and a high degree of incision, e.g. a scar as potential starting zone, increases the hazard. There is furthermore greater hazard at streams with north- to northeasterly aspect (cf. section 5.4). Even very gentle sections of a stream course far removed from a steep slope may be hazardous in extreme cases. The avalanches usually also affect a narrow zone bordering on the stream channels.

5.8.2 Meteorological conditions associated with increased hazard

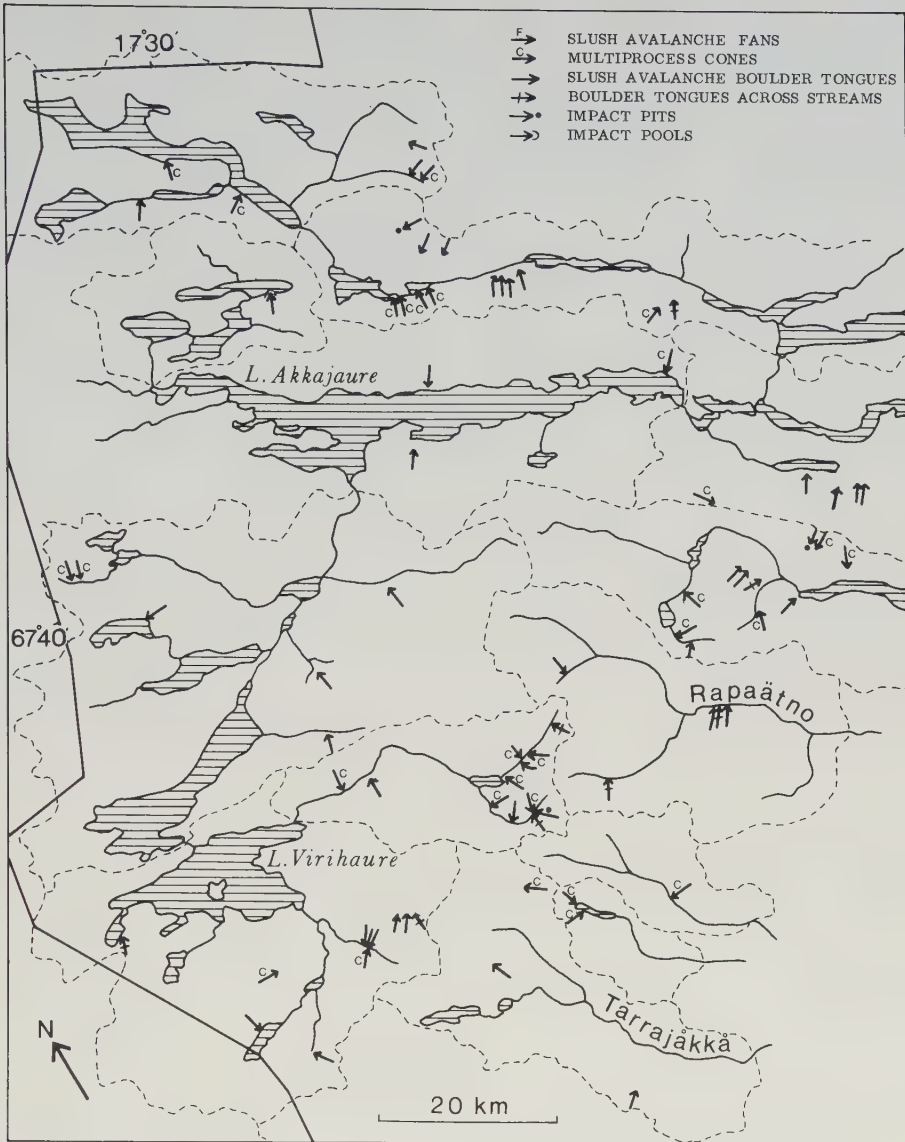
No more can be said on this topic with the present insufficient



a

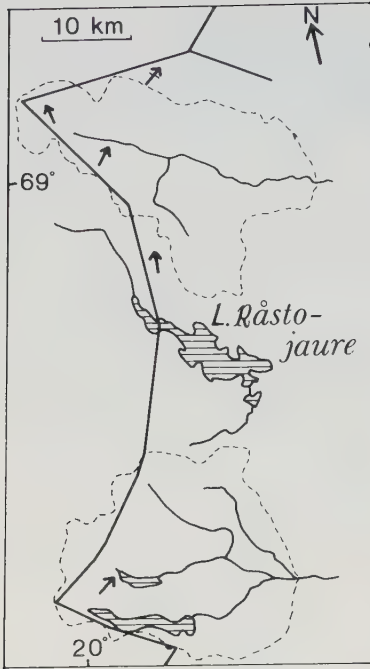
Fig. 5.48. Regional distribution of slush avalanche sites of geomorphological significance in northern Swedish Lappland (a-d). Morphological varieties concerning debris deposits or erosional features in the runout zones indicated by indices. Orientation of arrows shows aspect of paths. Major drainage basins are shown by dashed lines. Cf. Map III for an overview.

knowledge except that a rapid rise in temperature to well above 0 degrees, capable of producing large amounts of free water in the snow cover, increases the hazard. It appears that there is a short period of risk of slush avalanches each spring lasting approximately one or two weeks, after which continued warm weather will not lead to release of slush avalanches. Presumably, this is due to the development of efficient drainage channels under the snow cover or subaerially, eliminating the potential

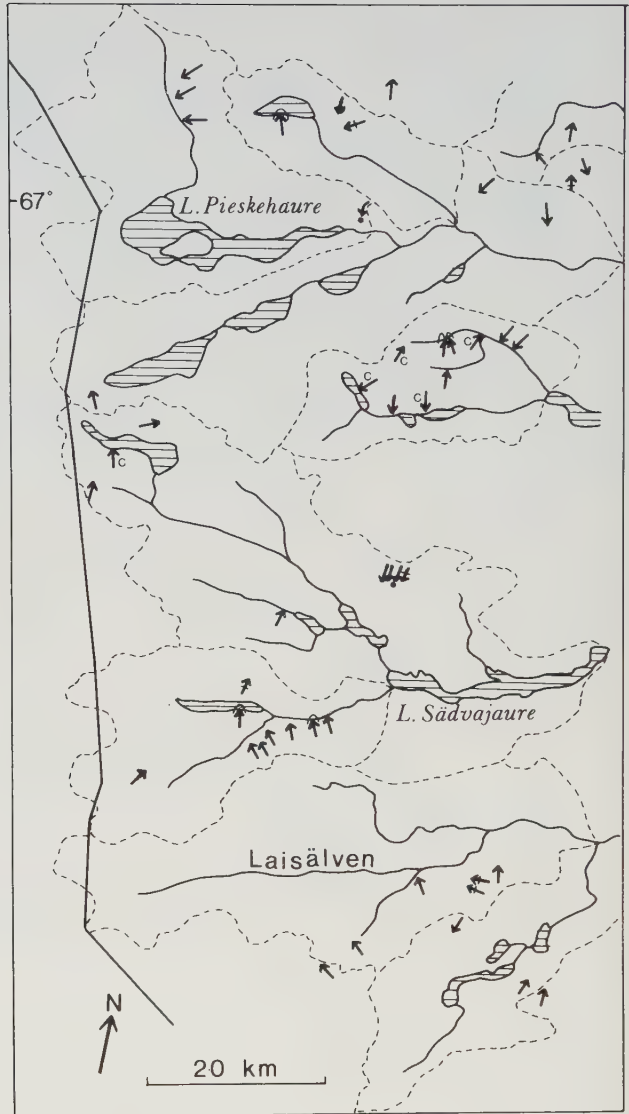


b

for melt water damming. Rapidity in melting as well as amount of snow and its characteristics, e.g. presence of thick basal ice crusts, probably determine whether a snow-covered stream will break up in a normal fashion or by slush avalanching. A useful monitoring programme with the aim of predicting the hazard would have to provide continuous recording of snow depth and free water content or melt water damming in a potential starting zone during the critical period in the spring. In areas with high avalanche frequency, known runout zones could be marked with warning signs on location, and on tourist maps.



c



d

5.9 Summary and conclusions

Slush avalanches occur with geomorphic effects along many first-order streams in northern Swedish Lappland. Compared with snow avalanches and debris flows, they have a more restricted distribution, but are not as strongly bound to the high mountain massifs with alpine relief. They may be released on slopes as gentle as 6° . Snow distribution in response to strong W-winds makes stream courses with aspects between N to E particularly prone to slush avalanching. The very long range of this type of avalanche makes it a potential hazard during the spring period, also on slopes of low gradient.

Two forms of release mechanism have been identified: 1) slab release of heavy wet snow resting on sloping bedrock ledges sometimes covered with ice crusts, and 2) water-saturation release of soaked slush, the motion of which may be triggered by a snow jam breakage, or possibly by the momentum of a flood released after temporary damming upstream, due to a snowbridge collapse. The release of slush avalanches generally occurs a few days after a marked temperature rise to above zero values, and is most typically associated with breakup of streams in spring. Individual sites seem to respond to local terrain and weather conditions in the timing of avalanche events. Widespread release of slush avalanches seems to occur only in years of exceptionally favourable combinations of snow conditions and rapid melting.

Observations in the Abisko mountains indicate a fairly high frequency of slush avalanches during the recent climatic conditions, contrary to earlier beliefs (Table 5.12). However, major deposits with old sub-relict distal parts, according to the vegetation cover, provide evidence for former events of larger magnitude, perhaps during the 'Little Ice Age'.

Erosion by slush avalanches may be high locally and result in long transport of material, but due to their comparative sparsity is of more limited importance on a catchment level. The local redistribution of slope debris results in stream channel scouring and deposition of debris tongues and fans at the foot of slopes or upon alluvial fans.

6. Observations on mass movements in Nissunvagge

6.1 Introduction

The data presented in the two previous chapters have provided some evidence of the geomorphological effects of debris flows and slush avalanches in northern Swedish Lapland. However, an assessment of the relative significance of the processes can only be made through comparison with the activity and occurrence of other geomorphological processes. In this chapter, an estimation of the importance of the rapid mass movements dealt with in the study in relation to other mass movements is attempted for Nissunvagge, using Map II as a basis. Two aspects will be treated, the spatial impact and the process rate or temporal frequency, the latter more tentatively as available data are limited.

6.2 Spatial impact of mass movements in Nissunvagge

The activity of periglacial mass movements, particularly those involving water, in fluid or frozen state, is generally seasonal in its character. Some of the processes are localized, such as rockfall, others are active over widespread areas, such as soil creep.

In Nissunvagge, spatial variations in process activity are governed by the large-scale valley morphology and its influence on slope gradient, local and microclimate, and by the distribution of different slope materials. Vegetation, itself dependent on these factors, is another agent influencing the slope processes. The previous or possibly present existence of permafrost on slopes may also be considered as important, as it affects the moisture conditions in surficial soil layers in spring.

A number of distinctive landforms are associated with the various processes. The more important of these forms of medium-scale are shown for central Nissunvagge on Map II, which also includes other features e.g. stemming from the deglaciation, often only slightly altered by postglacial processes. The distribution of various mass movements will next be treated in some detail.

6.2.1 Rapid mass movements

Rockfall is a widespread phenomenon in the valley, due to the high frequency of rockwalls (cf. Map II). There is a relationship between height of the walls and size of the corresponding talus accumulation (Figs. 6.1, 4.26). The importance of debris flows and snow avalanches in the reshaping of rockfall talus, especially in front of major rockwall chutes, was mentioned earlier (Section 4.3.6). Avalanche talus with well developed avalanche boulder tongues are prevalent on the western valley side (cf. Fig. 2.7), due to the importance of snow avalanches on slopes to the lee of strong winds, while debris flow cones dominate the eastern valley side (Fig. 2.5). True talus cones are therefore not well developed, and rockfall talus occurs mainly as 'simple

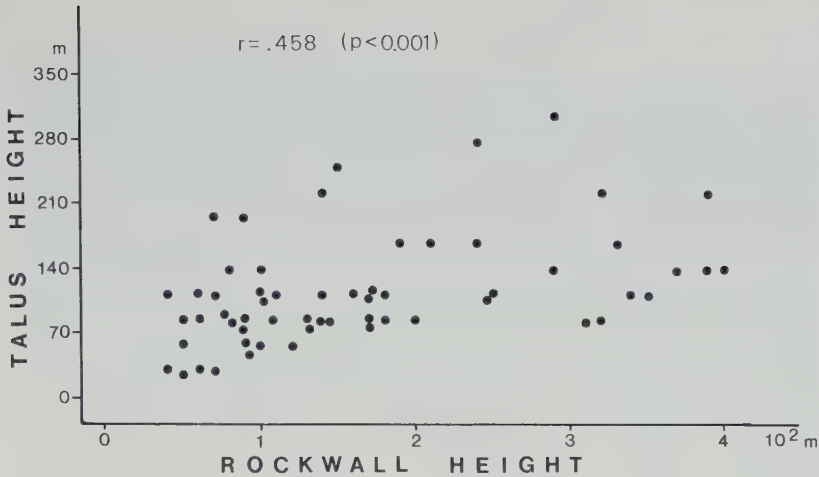


Fig. 6.1. Relationship between height of rockwall and talus in Nissunvagge. Based on measurements on Maps I and II, using a systematic sampling along the valley sides.

talus slopes' (Rapp 1959) beneath most rock walls. Debris flows also occur on simple talus slopes, with no clear connection to rockwall chutes. Minor flows (not shown on the maps) are localized to the steep slopes of the Nissunjokk canyon. They have contributed to the development of gullies or nivation hollows on the canyon sides.

Slush avalanches are evenly distributed between the valley sides and localized to small stream courses, some in valley bottom positions. They have the highest transporting power of the rapid mass movements, in several cases reaching the main stream of Nissunvagge (Fig. 6.2). The shortest debris transport by rapid mass movements is that due to small earth slides, released at the fronts of gelifluction lobes on gentle footslopes (cf. Fig. 4.4).

6.2.2 Slow mass movements

Evidence of slow mass movement on the slopes is widespread. Gelifluction lobes are frequent on the lower slopes on the western valley side (cf. Map II). Pattern ground features indicate smaller downslope movements on the opposite lower valley side, but strong frost action. At higher levels, slow mass movement is demonstrated by the occurrence of stone stripes (sorted stripes) in several parts of the blockfields, especially on the eastern side (Mt. Nissuntjärro). A part of the talus on this side appears also to have been affected by slow downslope movements, whereby basal block tongues of low inclination have developed (Fig. 6.3). However, the well-preserved state of the deglaciation features on the western valley side, close to the Nissunjokk canyon, shows that mass movement has been very limited in certain materials.



Fig. 6.2. The slush avalanche path along the tributary valley (section B, background) reaches the main stream in the valley bottom of Nissunvagge (foreground). The largest avalanches even cross the stream and deposit their debris load on the distal bank (cf. Map II). The stream channel downslope of the tributary valley was heavily reworked during the rainstorm in 1979. The site of the meteorological screen in Nissunvagge marked by an s. Gelifluction lobe for measurements of ground movements and temperatures indicated by an x. Air photo N.Å.Andersson 25.7.79.

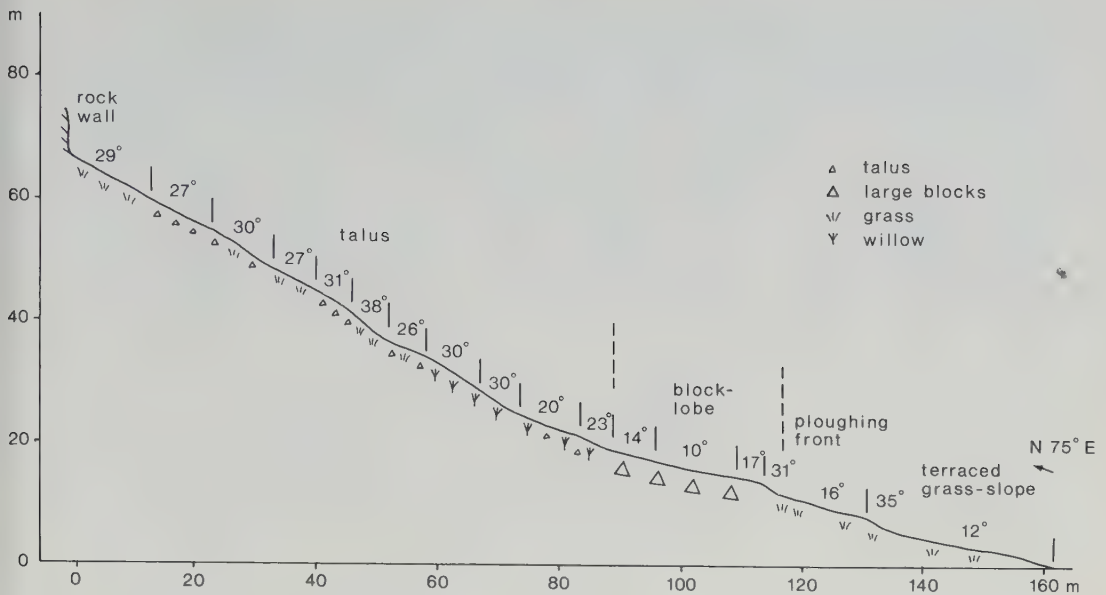


Fig. 6.3. Talus slope in northern part of section C, Nissunvagge, with basal block tongue of low inclination. The morphology may be due to slow mass movement of gelifluction type, possibly previously as a beginning rock glacier.

6.2.3 Spatial rankings

To be able to quantitatively compare the spatial impact of the different mass movements in Nissunvagge, a grid net was overlaid the map (cells corresponding to 200x200 m) and the presence or absence of the types of mass movement was noted for each square within different valley side sections. The significance of the processes in terms of occurrence in the different sections is shown by Fig. 6.4, and can be expressed by the ranking between the mass movements for each section. For the rapid mass movements only the resulting depositional landforms, or in the case of slush avalanches, runout zones, have been taken into account. The occurrence of rockwalls as source areas for rockfalls is also included in the Figure.

Debris flows were ranked first in sections c and d, and second in section b, all three the most heavily affected ones by the 1979 rainstorm. On the two sections that did not gain any new debris flows in 1979, i and k, they only ranked as number four. The localized nature of slush avalanching gave low rankings in most sections. The distribution of three types of mass movements appear to be aspect-related for the valley: snow avalanches with rankings 7-6-6 on the eastern side and 4-3-4 on the western; gelifluction with corresponding rankings 4-4-4 and 1-1-3; and blockfield creep (sorted stripes) with ranking 1-1-3 and 2-6-5, that is, preferred occurrence on the eastern side. The asymmetric distribution of snow avalanches and gelifluction lobes are

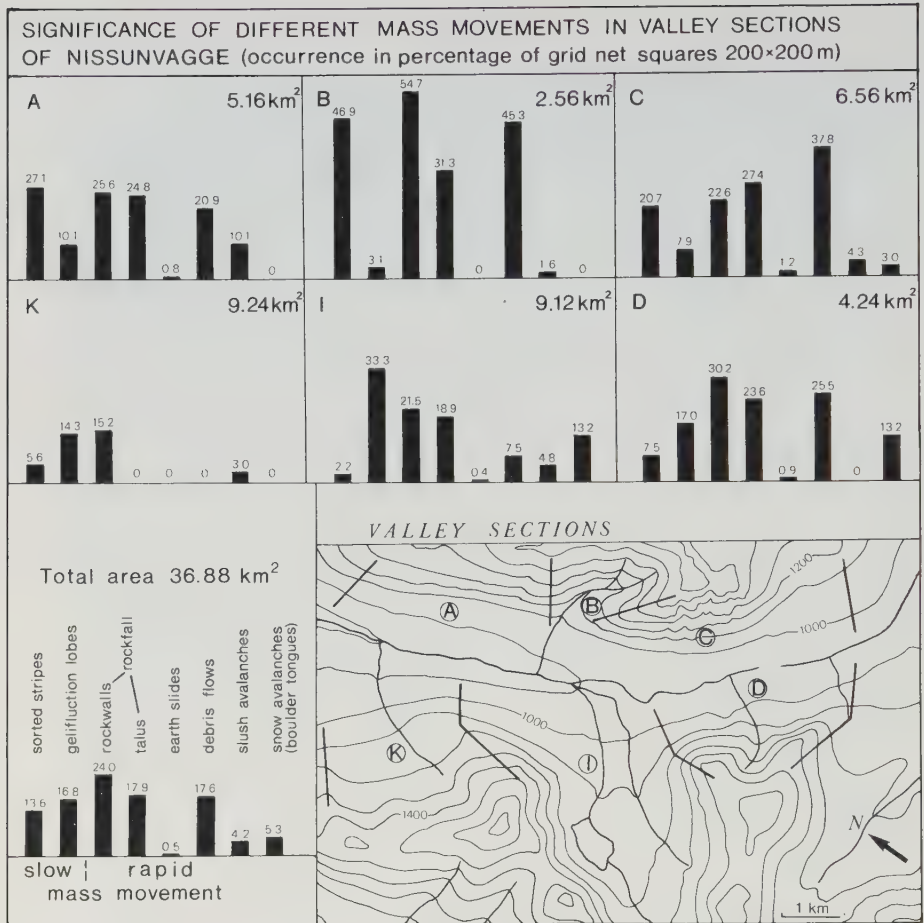


Fig. 6.4. Rankings of different mass movements on the basis of occurrence in valley sections of Nissunvagge. Cf. text.

probably both attributable to the snow accumulation pattern. This relationship is obvious for avalanches, but for gelifluction of indirect nature, reflecting wetter conditions caused by more long-lasting water supply compared with the opposite slope with less snow cover and stronger sun-exposure. No logical explanation for the observed distribution of sorted stripes can be offered presently.

For the whole area, rockfall was the areally dominating mass movement, if we exclude the ubiquitous creep in loose materials, followed by debris flows, gelifluction and blockfield creep/sorting. Avalanches of both types were areally taken rather unimportant, as was short earth slides.

6.3 Frequency or rates of processes

The available information on the recent activity of the different mass movements is incomplete. Only rough estimates can therefore be made on their relative importance. The frequency of debris flows and slush avalanches is assessed mainly from the licheno-metric data presented in section 4.4 and the information provided

in section 5.6. Measurements of gelifluction and creep rates in Nissunvagge were initiated during the study, but are of too short duration to give but indications of probable values. Data for snow avalanches and rockfalls are lacking. An extension of the initiated studies to comprise all types of mass movements and also other denudational processes in the valley would be valuable as an effort to obtain complete data on denudation for comparison with e.g. the Kärkevagge data (Rapp 1960b). For the present, the discussion must be restricted to the mass movements hitherto dealt with.

6.3.1 Gelifluction and creep

Measurements were initiated in 1980 and 1982. The installations comprise two lines of wooden pegs in gelifluction lobes on either side of the valley, two painted lines across debris flow lobes of 1979 in sections A and C, and one buried vertical profile of Rudberg pillars in a gelifluction lobe in section C (cf. Fig. 6.2). Air and ground temperatures at different depths were also measured in 1982 in one of the studied gelifluction lobes, using a data logger with one hour sampling interval. The purpose of the latter measurements was to assess the relationship between average yearly movement and number of frost cycles in the ground for various points of a gelifluction lobe. The results will be reported on in a separate paper.

The measurements of present rates of slow mass movement in Nissunvagge indicate the yearly averages shown in Table 6.1. These values refer to surficial movements. They may somewhat underestimate actual rates, as the fix points for the lines were not in bedrock, but on apparently stable large boulders. The test pillars for vertical movements have not been re-excavated as yet.

An attempt to estimate gelifluction movement on a longer time-scale was made by dating of buried organic horizons in gelifluction lobes that had been cut through by debris slides of 1979 (slide section C1-C5). The following dates were obtained (Håkansson 1984):

	distance from front of lobe, cm	depth beneath surface, cm	sample lab.no	C14-age years BP
lobe A	175	130	Lu-2161	310±45
			Lu-2161A	320±45
lobe B	160	75	Lu-2162	600±45

Due to the possible 'contamination' of the samples by irrelevant organic material, it is uncertain if the ages obtained can be used to infer time of burial and subsequent movement (cf. Everett 1967, Karlén 1982). Compared with the recent rates as indicated by the initiated measurements, the datings would imply slower long-term rates, or between 0.25-0.5 cm/year.

6.3.2 Debris flows and slush avalanches

As 'real-time', direct measurements are not advisable for these processes, an estimate of their geomorphic significance has to rely on available data on average amounts of erosion and frequency in time. Return periods for debris flow events in Nissunvagge are estimated on two different scales, related to the spatially varying impact of rainstorms (or snowmelt). Large events ('regions' or 'cells', cf. p.92) are assumed to recur with an interval of about 300 years, based on the lichenometric data.

Table 6.1 Measurements of gelifluction rates in Nissunvagge during the period 1980-1985 (except second site, 1982-1985).

site	average yearly movement (cm)	range in yearly movement (cm)
gelifluction lobe, W slope, opposite hut	1.0	0-2.3
gelifluction lobe, E slope, section C	4.5	1.8-7.8
blocky debris flow lobe, section A	0.3	0-2.3
sandy-stony debris flow lobe, section C	0.4	0-2.0

For the whole of northern Lapland return periods are shorter, even if the two recent cases at Tarfala in 1972 and Nissunvagge in 1979 represent an extremely close spacing in time. Isolated events ('spots'), including also very small features such as the ones released by the limited rainstorm of July, 1982 in Nissunvagge, are assumed to have recurrence intervals of 10 years, considering that triggering rainfall amounts, although intense, may be as low as 15-30 mm (section 4.3.8).

The frequency of slush avalanches has been estimated at 10 years from the evidence of recent activity (Table 5.12) and the stratigraphic data referred to in section 5.6.3. The accretion rate of at least 0.5 mm/year for depositional zones of slush avalanches is sufficient to account for observed landforms with estimated thicknesses of 4-5 m, built in postglacial time. Assuming an average amount of accretion of 5 mm for each avalanche gives return periods of 10 years during 9000 years. Possibly, these values are too high for Nissunvagge, where large landforms due to slush avalanches, comparable to cases in Kärkevagge and Kårsvagge, are missing. However, this may be related to removal of the material by the main stream of the valley.

6.3.3 Activity rankings

A comparison of the geomorphic activity in terms of mass transfer by debris flows and slush avalanches on the one hand, and gelifluction and short earth slides on the other is shown for Nissunvagge in Table 6.2. The vertical mass transfer of slope debris by the different processes has been calculated according to the formula $M = V \cdot \gamma \cdot (d \cdot \sin \alpha)$ where V = volume of debris (m^3), γ = its density (Mg/m^3), d = transport distance (m) and α = average slope angle. This estimate, although crude, indicates a higher long-term activity of the two major forms of rapid mass movements compared with gelifluction in the catchment. The data offer an opportunity to compare the estimated geomorphic significance of the processes dealt with for this valley of continentally influenced climate and amphibolite bedrock with the corresponding data from Kärkevagge, with its more maritime climatic influence and bedrock of mica-schists, cf. Table 6.3. The estimation was made that slush avalanches transport on the average 250 m^3 debris (175 m^3 rock volume) every 10 years in Kärkevagge, based on Rapp (1960b) and Table 5.12. The rapid mass movements were more important in relation to slow mass movements during the study period in Nissunvagge, compared with Kärkevagge. However, the extraordinary large magnitude of the debris flow event in 1979 in Nissunvagge has raised the numbers above their actual long-term

Table 6.2. Estimated denudation and debris transfer on slopes in Nissunvagge (sections A, B, C, D, I, K) during the period 1979-1985 and on a longer time-scale by selected mass movements.

	debris volume (m ³)	density	tons (rock volume)	tons/km ² a ⁻¹	average movement (m)	average gradient	ton-metres (vertical)	rock denu- dation (mm)	estim. return period	long-term rock denu- dation (B)
DEBRIS SLIDES										
AND FLOWS, multiple events (1979)	103 000	1.8	130 000	500	300	33	21.2 · 10 ⁶	1.9	300	5
isolated events (1982)	20	1.8	25	0.1	30	33	410	0.0003	10	0.03
SLUSH AVALANCHES ¹	500	2.6	910	3.5	800	15	1.9 · 10 ⁵	0.009	10	1.4
EARTH SLIDES	15	1.8	18.9	0.07	30	11	108	0.0003	10	0.03
GELIFLUCTION	1.55 · 10 ⁶	1.8	-	-	0.01	15	7 300	-	-	-

¹based on 3 active sites

Table 6.3. Estimated geomorphological importance of three types of mass movements on slopes in Nissunvagge 1979-85 and Kärkevagge 1952-60. Data for Kärkevagge by Rapp (1960b). RP=estimated return period in years.

	Nissunvagge	Kärkevagge
bedrock	amphibolite	mica-schists
altitudinal range (m)	700-1740	600-1590
area (km ²)	37	15
DEBRIS SLIDES AND FLOWS		
long-term rock denudation (B)	5 (RP 300)	1 (RP 200)
material transfer (ton-metres vertical a ⁻¹)	3 027 · 10 ³	96 · 10 ³
SLUSH AVALANCHES		
rock denudation ²	1.4 (RP 10)	3.5 (RP 10)
material transfer	26.9 · 10 ³	20 · 10 ³
GELIFLUCTION		
area on slopes (km ²)	6.2	2.2
average movement yearly (cm) ¹	1.0	2.0
material transfer	7.3 · 10 ³	5.3 · 10 ³

¹layer of 0.25 cm²based on 3 active sites

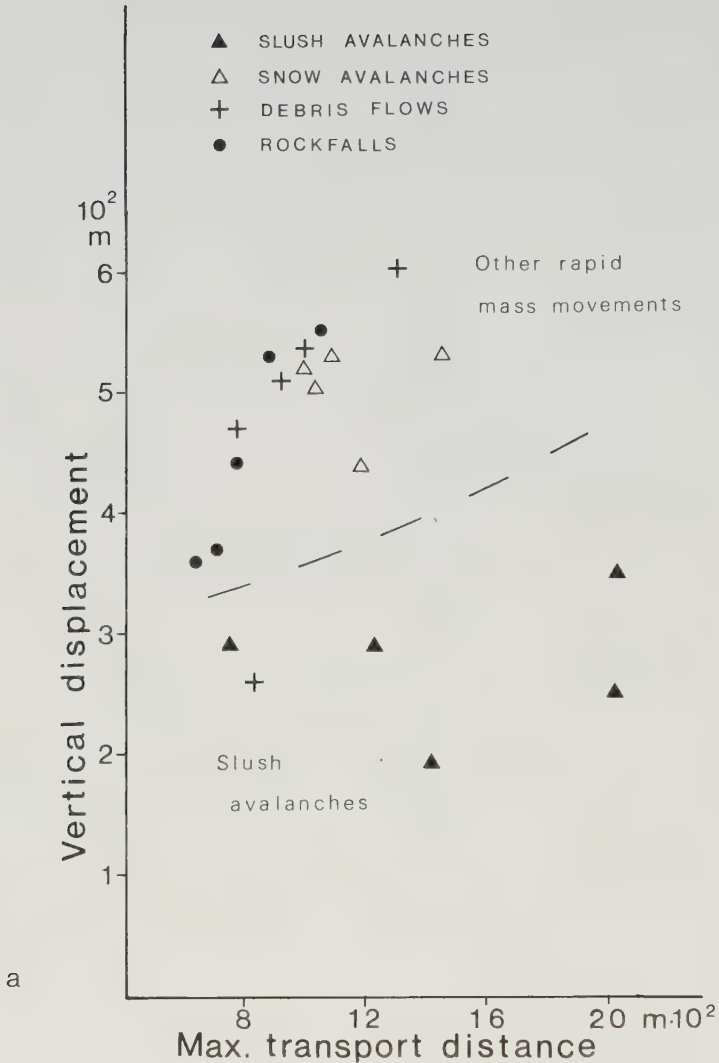
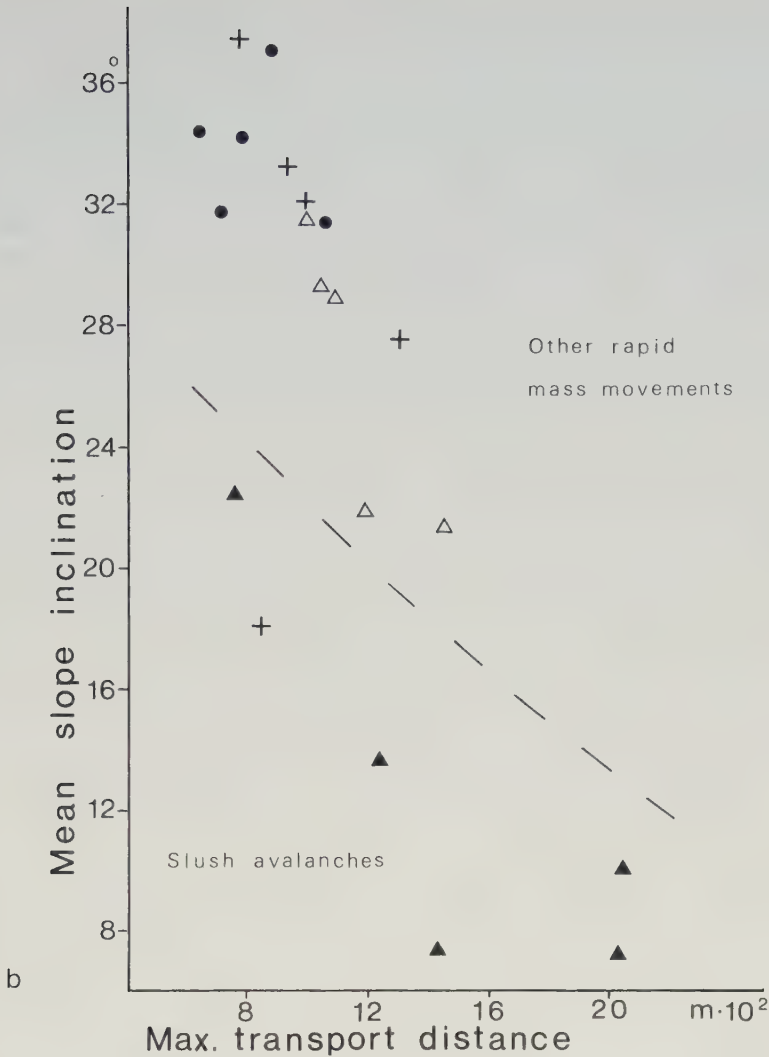


Fig. 6.5. a. plot of maximum transport distance of coarse debris vs. vertical displacement for various types of rapid mass movements b. plot of maximum transport distance of coarse debris vs. slope inclination for various types of rapid mass movements. A distinction can be made between slush avalanches and other processes in terms of transporting capacity. Based on data from Nissunvagge, measured on Maps I and II.

significance. Debris flows are therefore the dominant form of mass movement in Nissunvagge, whereas slush avalanches are comparatively more important in Kärkevagge. Gelifluction is of similar significance in both localities, according to the estimations.

The ranking of processes obtained may be compared with data from a catchment of somewhat smaller size, where rapid mass movements also exceed the slow ones in importance (Caine 1976, Barsch &



Caine 1984). Avalanches were found to be of lower significance than debris flows in that catchment. The activity rankings of Table 6.3 represent the amount of material eroded and transferred by the processes. A ranking may also be performed with respect to the transporting capacity of the mass movements (size and transport distance of material moved by the processes). This procedure reveals a marked contrast in terms of transporting capacity between slush avalanches and other rapid mass movements. The ability to transport coarse debris for long distances even across gentle slopes distinguishes slush avalanches from other forms of rapid mass movements (Fig. 6.5).

The approximate nature of the data presented should be noted, as the estimations both of return periods and average erosion are fairly rough. The importance of debris flows is probably overrated, as the magnitude of the 1979 event was most likely unusual in the postglacial time perspective. However, further field data are needed of the occurrence of old flows to corroborate this assumption.

6.4 Discussion and conclusions

Actual measurements or observations of the activity of the processes in Nissunvagge have given indications of present rates, which may more or less correctly be extrapolated backwards in time, depending on above all the character of former climates. The climatic factors may be considered as inherent in the frequency estimations made for debris flows (cf. Fig. 4.38), but are unrelated to presently measured rates of gelifluction. The landform record (distribution and characteristic data for various forms) provides an accumulated morphological response with respect to earlier process rates and characteristics. Additional stratigraphic data and information gained from datings of present or former vegetation have been used to complement the morphologic data in estimates of process rates for longer time periods.

The data gathered indicate a higher geomorphic importance of large-scale rapid mass movements with long return periods compared with slow continuously acting mass movements (Table 6.3). The overwhelming importance of widespread release of debris flows in terms of accomplished erosion is brought out very clearly from the Table. The study period coincided with an extreme event, raising the significance of debris flows with a factor of several hundred compared with their estimated normal annual material transport. The relative insignificance in the long run of smaller debris flow events even with fairly high frequency is a further conclusion that can be drawn from the data.

Slush avalanches, although more spatially restricted than debris flows in the valley, also accomplish considerable erosion according to the estimates.

The 1979 event in Nissunvagge offers a striking example of strong erosion within slope systems mainly leading to internal redistribution and continued storage of material, but relatively small output and hence denudation on the catchment level. Thus, despite the desirability pointed out by i.a. Caine (1976) to integrate work on hillslope erosion and catchment stream erosion, it is understandable why a separate treatment has often been followed. Only recently, attempts have been made to quantify the transfer of material between the mass wasting and fluvial transport systems in periglacial areas (Caine 1976, Strömquist 1983). The complexity of the matter is underlined by the possibility of identifying several subsystems within catchments, related to either spatial position (interfluve/ valley side slope/channel geometry subsystem, Chorley & Kennedy 1971) or nature and capacity of mass transport (solute/fine sediment/ coarse debris subsystem, Barsch & Caine 1984). Movement of coarse debris is clearly limited by slope gradient and therefore dependent on large temporary inputs of process energy to a higher degree than the continuously working processes of the solute and fine sediment subsystems (Op. cit). The 1979 event activated the coarse debris subsystem of the valley temporarily. For several years following this large energy input, slower readjustments on the slopes will occur, mainly manifested in increased flux of sediments through the fine sediment subsystem (higher turbidity in streams, in colloquial language).

In the Nissunjokk catchment a momentary erosion of about 1.9 mm occurred in the 1979 event due to debris flows. Assuming a recurrence interval of 300 years of similar major events, as calculated for the valley sections of Nissunvagge in Table 6.2, the average erosion rate by debris flows would be 5.8 (mm/1000 yrs). Thus, if these estimates are reasonably correct, the isolated event during a few hours time on June 23, 1979 accomplished more erosion than is normally attained in several hundred years. No value of total annual denudation is available for the valley. Studies in Kärkevagge have shown that chemical solution may be important (Rapp 1960b, Rehn 1985, Rehn in prep.), whereas slope wash appears to be of minor significance, except locally during the snowmelt period (Strömquist 1983, 1985). Disregarding our ignorance in these respects for Nissunvagge, the 1979 event must rank as a highly 'effective' geomorphic event in terms of landscape modification (Wolman & Gerson 1978).

7. Geomorphological significance of debris flows and slush avalanches

general conclusions

7.1 Introduction

As mentioned previously, the geomorphic importance of mass movements is a function of the magnitude of individual events and their frequency in time (Wolman & Miller 1960). These two aspects have been dealt with for debris flows and slush avalanches in chapters 4 and 5, and in comparison with slow mass movements in chapter 6. In this chapter, a synthesis will be attempted to assess the significance of the processes on a wider scale.

Catastrophic geomorphological events of the kind investigated influence slope development in a steplike manner, by the exceeding of strength-related thresholds in slope materials by external stresses (landforming event) and subsequent slower form adjustment until renewed incidence of extreme triggering conditions (Selby 1982 p. 219). The frequency of the landforming events varies in different climatic regions, giving contrasting rates in slope evolution. The downslope material transport by these processes provides a link between slope units otherwise largely developing individually by slower acting processes such as gelifluction and surface wash (Strömquist 1983).

The importance of erosional events can either be assessed from their amount of accomplished denudation, or with respect to their modifying influence on slope morphology (denudation vs. relief-forming effects, Starkel 1976). The persistence of the slope modifications is dependent on the normal rate of change or denudation, which varies for different climatic zones. By comparing erosional events to normal annual rates a measure of the effectiveness of the event can be obtained (cf. Wolman & Gerson 1978).

7.2 Regional geomorphological significance of debris flows and slush avalanches

An assessment of the geomorphic activity of debris flows and slush avalanches on the regional scale can only be made tentatively at the present state of knowledge, and mainly qualitatively. The following discussion is primarily included as an attempt to relate the estimated significance of the processes on the local scale of Nissunvagge and Kärkevagge to their probable large-scale importance, on the basis of the data concerning regional distributions presented earlier.

It is evident that the average densities of both slush avalanches (about $0.15/\text{km}^2$) and debris flows (about $7/\text{km}^2$) in Nissunvagge greatly exceed the regional values. These are approximately known for slush avalanches (about $0.75/100 \text{ km}^2$), but more difficult to estimate for debris flows due to the more generalised knowledge of their distribution (cf. section 4.5). For slush avalanches,

extrapolation of the observed erosion and estimated frequency at sites in Nissunvagge (250 m³ and 10 years) to 225 sites in the total study area (30 000 km²) yields a regional value of material transport corresponding to 0.13 B, thus about ten times smaller than the 1.4 B obtained for Nissunvagge. The value is derived only to give an idea of the more limited significance of slush avalanches on the broader scale, due to their relative low spatial frequency in the area.

A corresponding estimate for debris flows cannot be made, although it may with reasonable confidence be concluded that their importance regionally is lower than in Nissunvagge, although probably higher than what is estimated for slush avalanches, in view of their more widespread occurrence. Whether the higher geomorphic importance of the treated rapid mass movements compared with gelifluction as indicated from the local data is valid also regionally is thus less obvious, at least concerning slush avalanches.

7.3 The problem of climatically induced variations in mass movement activity

Generally, rapid mass movements of the debris slide and flow type are primarily linked with steep relief and strength thresholds of the slope materials, rather than any particular climatic region (Carson 1976). Snow and slush avalanches are naturally confined to cold environments with abundant solid precipitation. However, as liquid water plays a significant role in the mobilization of both debris flows and slush avalanches, and its availability is governed by meteorological factors, any climatic variation affecting these factors might influence the frequency and/or magnitude of the processes. Such possible influence is a complicating factor in any attempt to draw conclusions on previous long-term activity of mass movements on the basis of studies of the recent activity of the processes.

In the case of debris flows, a changed incidence of rainstorms with generally wetter conditions may affect slope hydrology, with increased pore pressures and liability of slide initiation. Thus a certain relation between climate and threshold slope angles is possible (Carson 1976). The relationship between previous debris flow frequency and climate was briefly treated in chapter 4.4, and it was concluded that available data are insufficient to allow any correlations at present.

The problem of climatic variations influencing the activity of mass movement processes is too complex to evaluate within the context of the present study. Certainly more needs to be learnt about the recent climatic controls on mass movement activity, before it will be possible to correlate previous climates and process rates in more detail. However, the field holds promise of interesting results with respect both to geomorphological and climatological angles of approach (cf. e.g. Grove 1972).

Although occurring sporadically and sometimes with long time interval, debris flows and slush avalanches are clearly part of the presently active morphogenesis in northern Lappland, having a strong influence on the evolution of many regolith-covered slopes. Although their frequency and importance were presumably even higher in late-glacial and early postglacial time, the landforms they have produced should not be viewed as primarily relict features in the landscape, but are in many cases still actively formed.

7.4 Conclusions on the geomorphological role of debris flows and slush avalanches in the study area

The geomorphological role of debris flows and slush avalanches in Swedish Lapland has been dealt with under two main aspects: the spatial impact (distribution) and morphological results of the processes on the one hand, and the present and former (postglacial) activity of the processes on the other. The conclusions reached concerning each aspect will be summarized point-wise below.

7.4.1 Distribution and types of process-related landforms.

1. The distribution of debris flow and slush avalanche landforms is the result of an interaction of two main sets of factors, which can be broadly referred to as slope characteristics (gradient, morphology, material, hydrology, vegetation) and climate characteristics (precipitation, wind, temperature). The great variability in these factors leads to an irregular occurrence of the forms, with a tendency to clustering for debris flows, which can be released in locally high densities following heavy concentrated rainfalls.

2. Both types of landforms are primarily restricted to areas above the tree-line in the high mountain area of Lapland. Debris flows in particular are strongly concentrated within high relief areas with steep slopes. However, a limited landforming influence by the processes occurs also within the mountain birch zone.

3. Spatially, debris flows have their greatest impact on talus slopes, where they contribute to the development of an increased basal concavity by deposition of material as typical landforms of medium-scale (debris flow levees, lobes and cones). Slush avalanches are geomorphologically important mainly along first-order stream courses, on which they impose a characteristic suite of small-scale landform features (erosional grooves, debris coatings, stripped surfaces exposing bedrock) and medium-scale terminal deposits (avalanche fans, boulder tongues) or erosional forms (impact pits and pools). Frequently the effect of slush avalanches is to reshape alluvial cones substantially, and a transitional landform between the latter and the slush avalanche fans or boulder tongues can be distinguished (multiprocess cones). Both debris flows and slush avalanches primarily act as within-basin transportational agents, affecting the coarse debris subsystem.

7.4.2 Recent and former process activity (magnitude and frequency)

1. The geomorphological capacity of individual debris flows vary between wide limits, or from a few m^3 to more than 10 000 m^3 . Slush avalanches may carry debris loads of several hundred m^3 , but are especially remarkable in their capacity to transport huge boulders across gentle slopes, with a longer range than debris flows in general. In this respect they are also clearly distinguishable from snow avalanches.

2. Morphological evidence and tentative datings by lichenometry indicate long return periods for debris flow events of major size, or about 200-300 years in the Nissunvagge area. Small debris flows may recur much more frequently, but have insignifi-

cant geomorphological impact compared to the large events of multiple release ('cells' or 'regions' of rainfall impact causing slope failures). Very old flows from the deglaciation period have not been identified and the present data are in favour of a continued debris flow activity throughout the Holocene, and in recent time, as witnessed by the events described in the study. Correlations between flow activity and climatic variations are for the present very uncertain.

3. Slush avalanches are active in the present climate on several sites in the Abisko mountains with recurrence intervals varying from a few years to several decades for geomorphologically significant cases. However, the vegetation cover (grasses, lichens, occasionally willows and birch) on many large deposits indicates earlier larger magnitude of events, or a very long return period for extreme events in terms of magnitude. The data from Nissunvagge (Table 6.2) and the regional survey of geomorphologically important sites (Map III) indicate a somewhat lower status of slush avalanches compared with debris flows with respect to denudation.

4. Indications from measurements in Nissunvagge support the results from other studies (Rapp 1960b, Caine 1976) of a higher geomorphological significance of the rapid mass movements dealt with compared with slow mass movements and thus suggest that slope development is partly related to the episodic crossing of thresholds of energy input to the geomorphic system. The effects of continuously working processes other than slow mass movements, have not been dealt with in the present study.

8. Summary

The thesis deals with two forms of rapid mass movements characteristic of the periglacial mountain environment of northern Swedish Lapland, with an emphasis on distribution and description of resulting landforms, and on the activity of the processes with respect to erosion and slope development. The study is concerned with debris flows and slush avalanches on two spatial scales: the local, exemplified by mapping and field study of the forms occurring in the Abisko area and in particular Nissunvagge valley; the regional, comprising analyses of the distribution of the forms in northern Swedish Lapland on the basis of available data and a survey on air photos of slush avalanche sites with geomorphological importance.

After an overview of the theoretical background to the study in chapter 1, a short description is given of the study area concerning its general environmental characteristics, on the local and regional scales (chapter 2). A presentation of the methods used in the study makes up chapter 3. Debris flows and slush avalanches are dealt with separately in chapters 4 and 5. The processes are first described regarding type of movement, initiating factors and terrain localization. Their erosional effects and the resulting landforms are then treated.

In the study area, debris flows occur as valley-confined flows or hillslope flows, the latter being the most common type. They play an important part in the total transport of weathered material by various slope processes. Slopes beneath rock walls are characteristically influenced by the activity of debris flows with respect to morphology of both detail forms (flow channels, block levees and lobes) and overall profile (basal concavity and cone development).

Slush avalanches were found to be mainly confined to minor stream courses, where they occur in years with catastrophic stream breakup in spring. Their geomorphological effects are scouring of the stream channels and accumulation of debris deposits, which may interfere with the streamflow and cause lateral shifting of the channel, as well as increased sediment transport downstream of a recent avalanche deposit.

Transporting capacity (magnitude) varies widely for both debris flows and slush avalanches, with a larger range for the former. Individual debris flows may carry 10 000 m³ of debris, or more. In terms of denudation, estimates for the Nissunvagge area indicate a larger significance of debris flows compared with slush avalanches, or 5 vs. 1.4 B (mm/1000 yrs). To obtain data on long-term significance of the processes, three dating techniques were applied: C14-dating, dendrochronology and lichenometry. Also some information of stratigraphic nature was utilized. The data indicate long recurrence intervals for major events of geomorphic importance, amounting to several hundred years. Smaller events may occur with only a few years interval within limited areas.

A special study was made concerning the release conditions of slush avalanches in the Abisko area. It was concluded that two forms of release are likely, as snow slab avalanches subsequently transformed into slush in a stream, or as mobilized water-saturated snow resulting from rapid melting and impeded meltwater drainage in the snow cover of a stream course.

The distributions of debris flows and slush avalanches in northern Swedish Lapland differ in several respects. Debris flows are more strongly bound to areas of steep relief, whereas slush avalanches also occur in areas with fairly gentle slopes. Debris flows often occur in clusters or series along valley sides, contrary to slush avalanches, which generally occur isolated. The difference is related to the significance of localized rainstorms of high intensity as triggering factors for debris flows, rather than rapid snowmelt, which is decisive for the release of slush avalanches. Debris flows and slush avalanches constitute natural hazards in the Swedish mountain area. Short descriptions of hazardous terrain sites and triggering meteorological conditions are given for both processes.

In the following two chapters (6 and 7), debris flows and slush avalanches are compared in terms of geomorphological significance with slow mass movements, which are being monitored in the Nissunvagge area. In agreement with results from studies elsewhere (Rapp 1960b, Caine 1976) the present data are interpreted in favour of a larger importance of the rapid mass movements in this area. It is concluded that the processes are still active and do not appear to have been mainly of importance for a short period after the deglaciation.

Appendices

APPENDIX 1. Notes on the vegetation in a small catchment, Mt. Nissuntjärro (section B).

The following list includes plant species found on the opposite valley sides (north- and south-facing) of section B in connection with the study of mass movements. The species are grouped according to occurrence (cf. Fig. 4.17), with no regard to frequency or claim of completeness for the valley as a whole, the data being sampled near microclimatic measurement points (Fig. 4.35). *Luzula* and *Carex* are not distinguished into species. The field work was carried out by M.Sc. Ann-Margreth Berggren, Lund.

1. Species with occurrence only on the north-facing slope

	(Swedish)
<i>Cardamine bellidifolia</i>	fjällbräsmå
<i>Pedicularis hirsuta</i>	fjällspira
<i>Ranunculus nivalis</i>	fjäll-smörblomma
<i>Saxifraga rivularis</i>	snöbräcka
<i>Saxifraga stellaris</i>	stjärnbräcka

2. Species with occurrence on both slopes

A. more observations on the north-facing slope

<i>Cassiope tetragona</i>	kantljung
<i>Diapensia lapponica</i>	fjällgröna
<i>Luzula</i> sp.	
<i>Oxyria digyna</i>	fjällsyra
<i>Ranunculus glacialis</i>	isranunkel
<i>Saxifraga oppositifolia</i>	purpurbräcka

B. similar occurrence on both slopes

<i>Carex</i> sp.	
<i>Cassiope hypnoides</i>	mossljung
<i>Festuca ovina</i>	fårsvingel
<i>Polygonum viviparum</i>	ormrot
<i>Salix herbacea</i>	dvärgvide
<i>Salix reticulata</i>	nätvide
<i>Saxifraga nivalis</i>	fjällbräcka

C. more observations on the south-facing slope

<i>Astragalus alpinus</i>	fjällvedel
<i>Dryas octopetala</i>	fjällsippa
<i>Draba nivalis</i>	isdraba
<i>Saussurea alpina</i>	fjällskära
<i>Saxifraga cernua</i>	knoppbräcka
<i>Sedum rosea</i>	rosenrot
<i>Silene acaulis</i>	fjällglim

3. Species with occurrence only on the south-facing slope

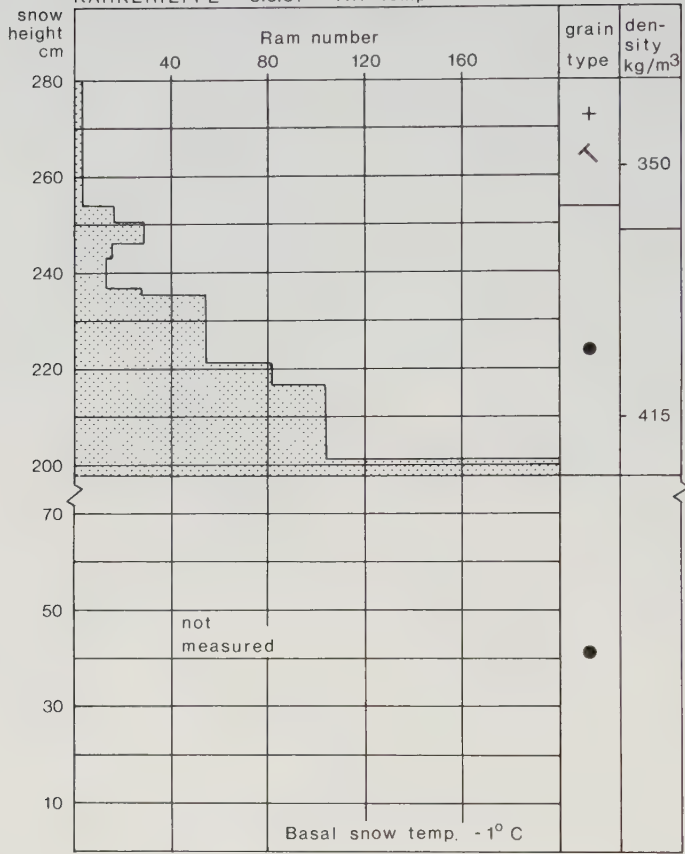
<i>Antennaria alpina</i>	fjällkattfot
<i>Arabis alpina</i>	fjälltrav
<i>Bartsia alpina</i>	svarthö
<i>Cerastium alpinum</i>	fjällarv
<i>Chamaenerion angustifolium</i>	mjölkört
<i>Erigeron uniflorus</i>	fjällbinka
<i>Gentiana nivalis</i>	fjällgentiana
<i>Hieracium alpinum</i>	fjällfibbla
<i>Parnassia palustris</i>	slätterblomma
<i>Phyllodoce caerulea</i>	lappljung
<i>Potentilla crantzii</i>	vårfingerört
<i>Roegneria borealis</i>	fjällem
<i>Salix</i> sp. (low shrubs, leaves pubescent)	
<i>Saxifraga cespitosa</i>	tuvbräcka
<i>Solidago virgaurea</i>	gullris
<i>Thalictrum alpinum</i>	fjällruta
<i>Trollius europaeus</i>	smörbollar
<i>Vaccinium vitis-idaea</i>	lingon
<i>Vaccinium uliginosum</i>	odon
<i>Veronica alpina</i>	fjällveronica
<i>Viola biflora</i>	fjällviol
<i>Viscaria alpina</i>	fjällnejlika

APPENDIX 2. Snow profiles and Ram penetration data from a slush avalanche site in Kärkevagge.

The data were sampled in the starting zone of slush avalanches at the Kärkerieppe cirque, Kärkevagge, cf. section 5.3.5 A. Key to snow grain type symbols (based on Perla & Martinelli 1975, p. 217):

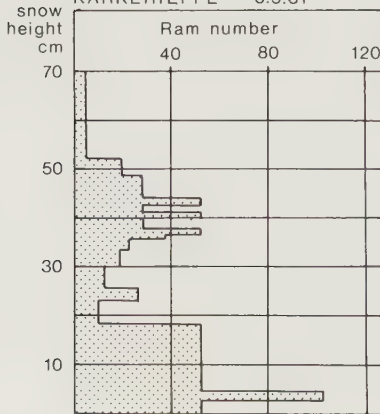
- + Freshly deposited snow.
- < Irregular grains, mostly rounded but often branched. Early stage of ET-metamorphism.
- Rounded, small grains (less than 2 mm). End stages of ET metamorphism.
- Rounded grains formed by MF metamorphism. Size usually more than 1 mm.
- ◻ Angular grains with flat faces. Early stage of TG metamorphism.
- ^ Angular grains with stepped faces, some hollow cups. Advanced stage of TG metamorphism.
- Ice layer.

Profile A

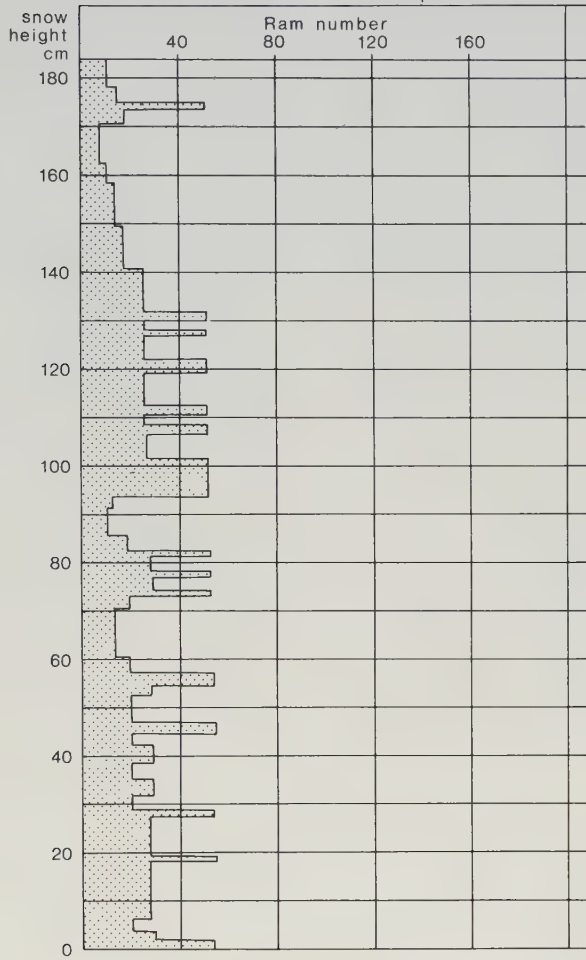
KÄRKERIEPPE 5.5.81 Air temp. -5°C 

Profile C

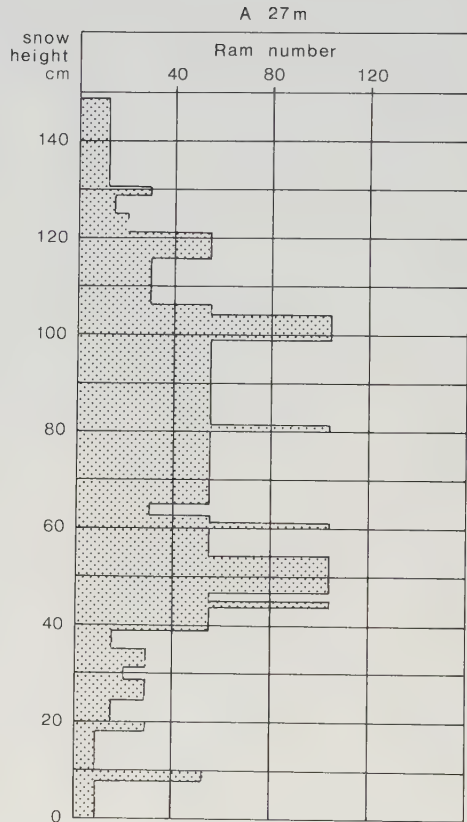
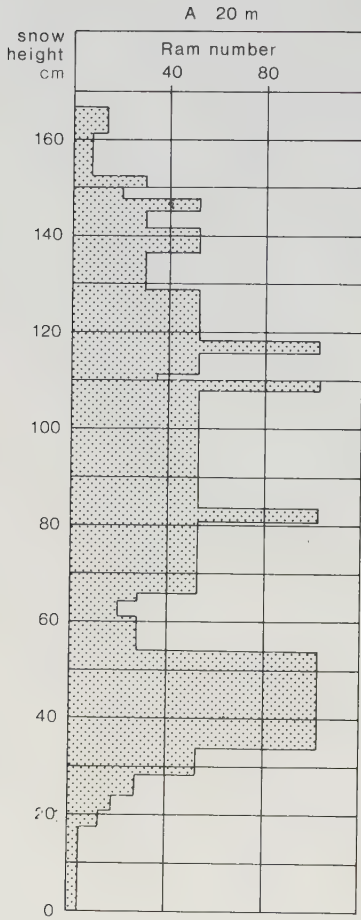
KÄRKERIEPPE 5.5.81



Profile 4

KÄRKERIEPPE 6.6.81 Air temp. -1°C 

30.4.82

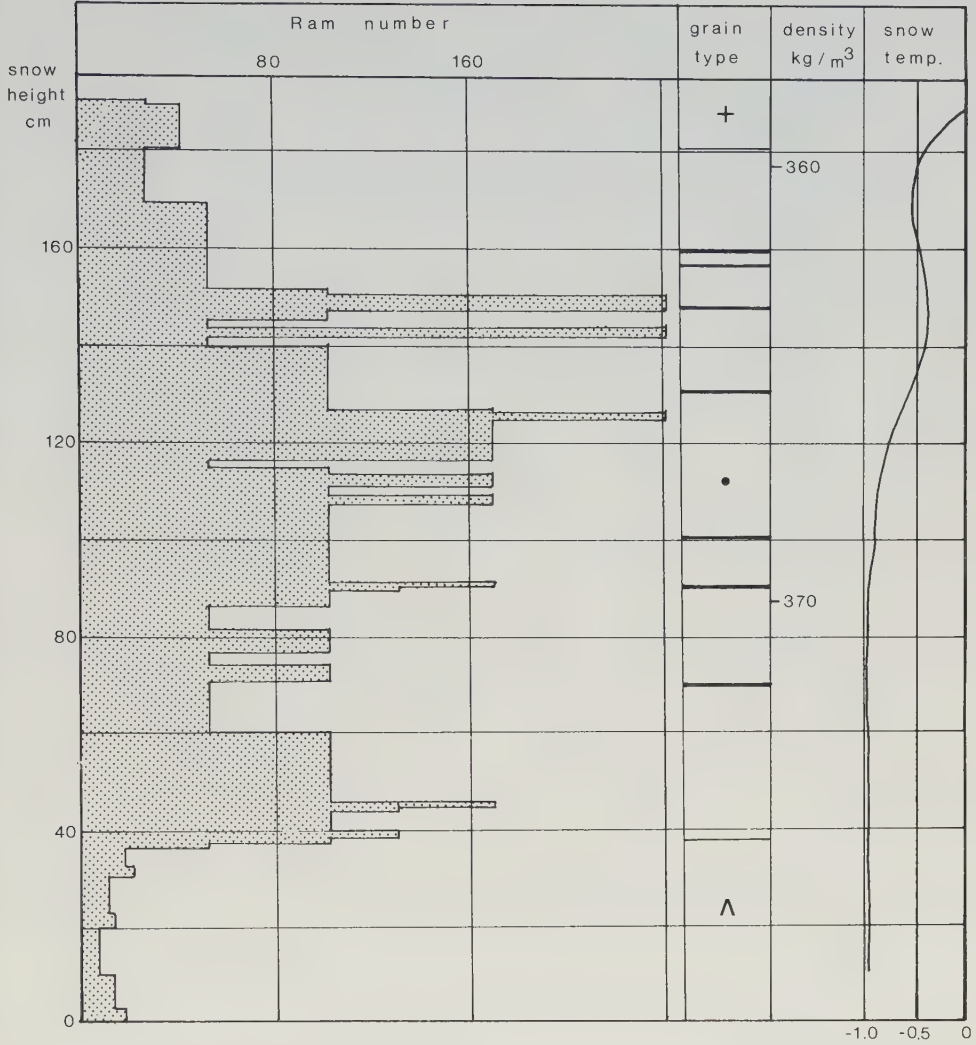


KÄRKERIEPPE

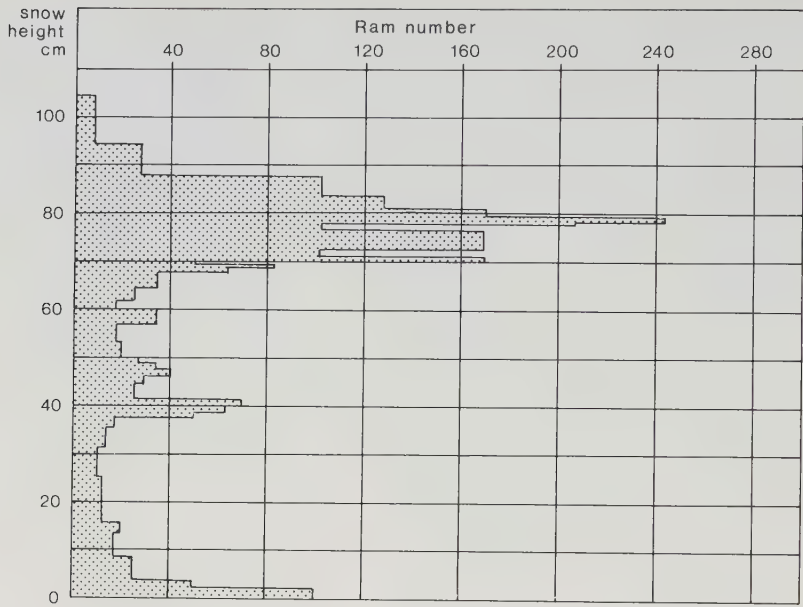
24.5.82

Profile A1

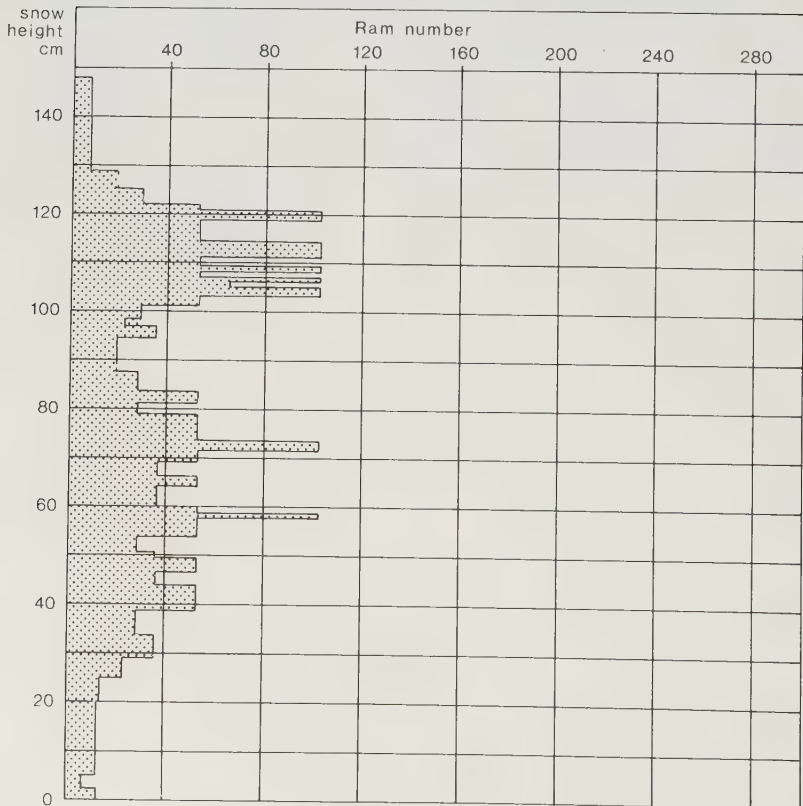
Air temp. +4.0



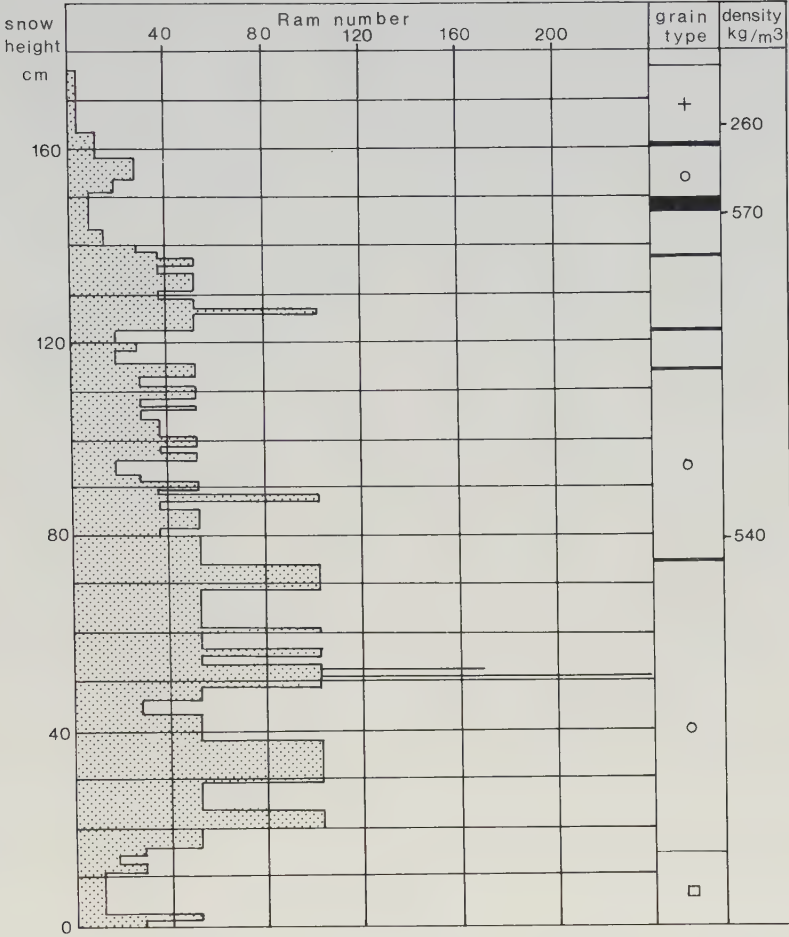
KÄRKERIEPPE 13.6.82 cross profile 10 m



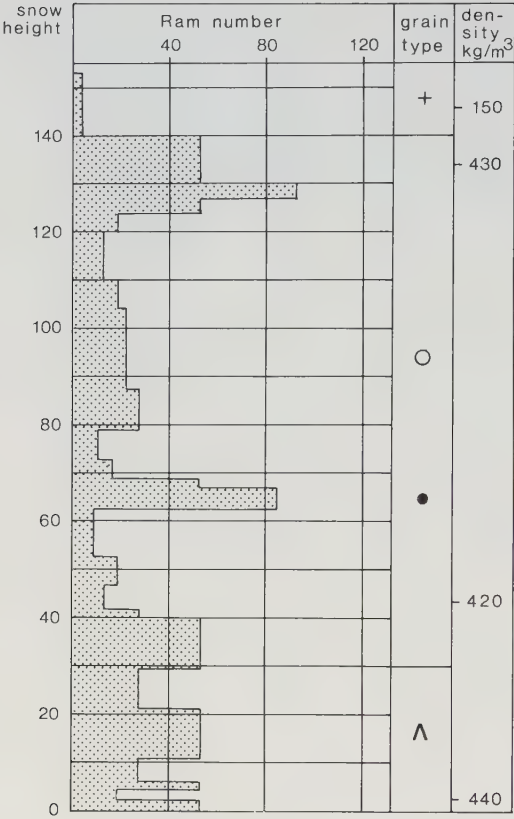
KÄRKERIEPPE 13.6.82 cross profile 30 m



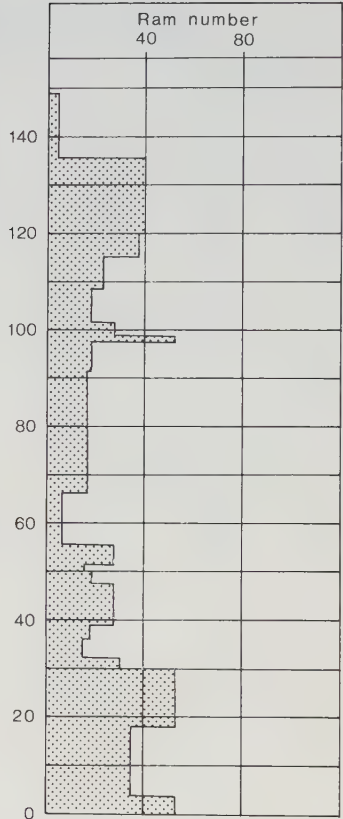
KÄRKERIEPPE 20.6.82 profile A1 Air temp. +0.5



KÄRKERIEPPE 5.5 83 Profile A1
Air temp. -3°C

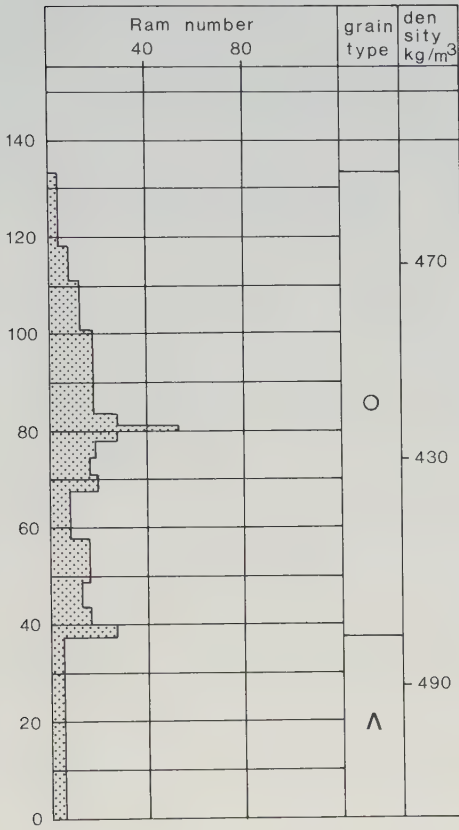


Profile A2
5.5.83



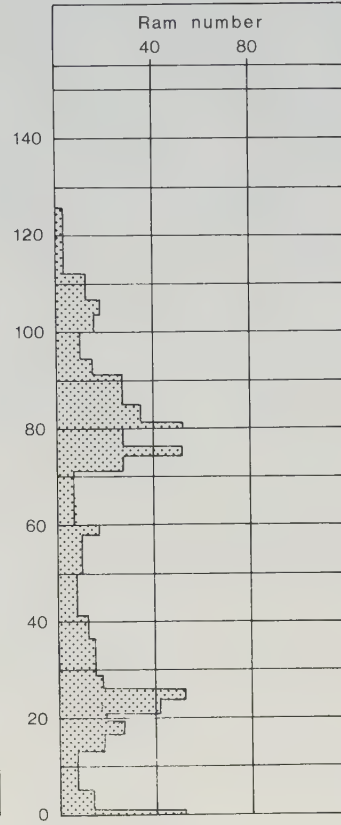
Profile A2

10.5.83

Air temp. $\pm 0.5^{\circ}\text{C}$ 

Profile A1

10.5.83



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